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SEISMOLOGICAL EVIDENCE FOR ROOTS OF MOUNTAINS

BY BENO GUTENBERG

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ABSTRACT

Theoretical problems concerned with the calculation of the velocities of seismic waves in the various layers of the earth's crust, and the determination of the thickness of such layers, are discussed. The results are applied to seismograms obtained from artificial explosions, and sources of error in interpretation are investigated. Seismograms registered in different parts of the earth from epicenters less than 1000 km from stations are used to investigate the layering in these regions and the problem of mountain roots. The bearing of the findings on the theory of isostasy is discussed.

INTRODUCTION

The hypothesis that mountains extend downward as well as upward was suggested by Airy (1855). He compared the state of the earth's crust "to the state of a raft of timber floating upon the water; if we

remark one log whose upper surface floats much higher than the upper surfaces of the others, we are certain that its lower surface lies deeper in the water than the lower surfaces of the others." This hypothesis of the "roots of mountains" plays an important role in the theory of isostasy. Geological, geodetic, and seismological findings support the mountain-root hypothesis. (*See,* for example, Daly, 1940, p. 415.) In the present paper seismological evidence pertaining to this problem is presented.

THEORY

To obtain quantitative results, we suppose that the earth's crust consists of a number of parallel concentric layers (Fig. 1b [1]) in each of which the velocity v is constant and higher than in that next closer to the surface. The source of an earthquake consisting of a single impulse is supposed at point H (depth h below the surface); E is the epicenter vertically above H. Under these assumptions, one group of rays travels along straight lines a in the upper layer. The impulses following these paths are recorded at the surface of the earth with rather large amplitudes (waves marked $\overline{P}$ in Plates 1 and 2). Near the epicenter E, they arrive almost simultaneously, as the distances HE and HA do not differ much. At greater distances, the beginnings of this phase (Fig. 1a) lie approximately along a straight line a; if this line is extended to the left, it intersects the time axis near the origin time O of the shock.

A second group of waves proceeds downward from the source H (Fig. 1b) at an angle of incidence α (between the ray and the vertical) so that $\sin \alpha$ is slightly smaller than the ratio v_1/v_2 of the velocities in the two layers. These rays are refracted into the second layer near J, later are refracted back into the first layer and follow the paths marked b. The impulses produced at the surface by these waves are smaller than those from the first group (P^* in Plates 1 and 2). Their arrival times lie along a straight line b (Fig. 1a). There is a similar group of rays c refracted near K, departing downward from H at an angle of incidence β for which $\sin \beta$ is slightly smaller than v_1/v_3. Their arrival times form the straight line c in Figure 1a. The times τ_2 and τ_3 (Fig. 1a) are called "intercept times", following Byerly.

At considerable distances, the differences in length of the rays a approach the length of the corresponding differences in distance near the surface; for example, $(HG - HF)$ is close to $(EG - EF)$. For most shocks used in this paper, the time error introduced by assuming that both are equal is less than 1 second at distances beyond 50 km. It is

[1] Illustrations in this paper were drafted by Mr. John M. Nordquist.

reduced by the effect of the curvature of the earth. At distances exceeding a few times, say five, the depth of the focus *(HE)*, the apparent velocity with which the impulse *a* travels along the surface, is approxi-

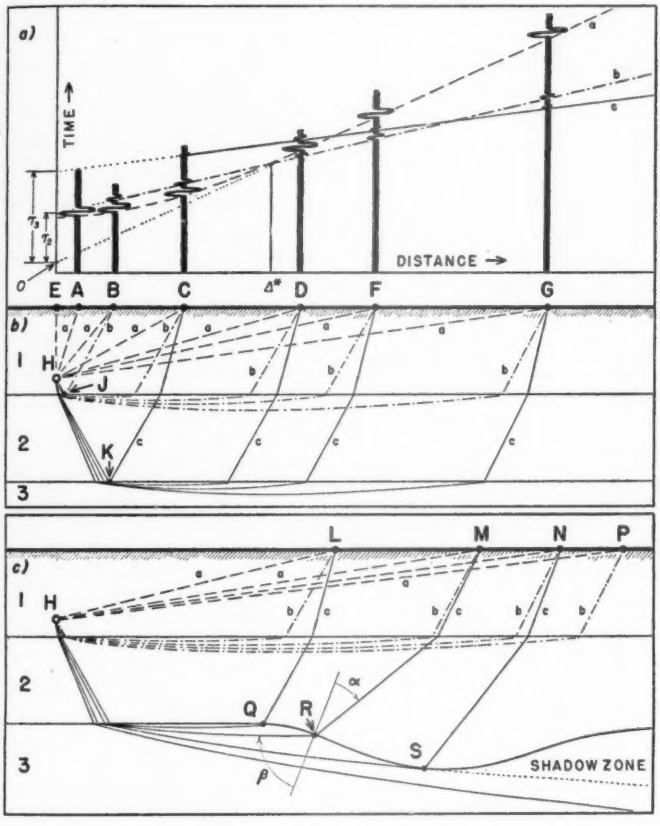


FIGURE 1.—*Sketch of wave paths, seismograms, and travel times*

mately equal to the true velocity v_1 in the first layer; near the epicenter this apparent velocity is greater. Similarly, the slope of lines *b* and *c* in Figure 1a is nearly equal to the true velocities v_2 and v_3 respectively; as the radii of the surfaces between the layers decrease with increasing depth, the apparent velocity found from the slopes of *b* or *c* in Figure 1a must be multiplied by $(r - d)/r$ (r = radius of the earth, d = depth

of the surface along which b and c respectively travel) to get the true velocities v_1, v_2, and so forth, in the various layers. Using Snell's law and neglecting the small effect of the curvature of the earth, the angles of incidence (i_1, i_2, i_3 . . . in the respective layers) of a ray having its deepest section in the n^{th} layer are given by

$$\sin i_1 : \sin i_2 : \sin i_3 : \ldots : 1 = v_1 : v_2 : v_3 : \ldots : v_n \tag{1}$$

Under the assumptions mentioned previously, neglecting the curvature of the earth and the increase of velocity with depth within each layer, it is possible to find travel times and corresponding distances simply by using trigonometric equations. For the following we assume, besides, that the source is in the first layer and we neglect the effect of sediments, which can be estimated if necessary. The case of a source in a deeper layer can be treated similarly but is of no interest in the present paper. If we have n layers, the thickness d_{n-1} of the layer $n-1$ can be found from

$$d_{n-1} = \frac{v_n\tau_n - (2d_1 - h)\sqrt{\left(\frac{v_n}{v_1}\right)^2 - 1} - 2d_2\sqrt{\left(\frac{v_n}{v_2}\right)^2 - 1} - \ldots - 2d_{n-2}\sqrt{\left(\frac{v_n}{v_{n-2}}\right)^2 - 1}}{2\sqrt{\left(\frac{v_n}{v_{n-1}}\right)^2 - 1}}, \tag{2}$$

where v_n is the velocity in the n^{th} layer, τ_n the intercept time (Fig. 1a), h the depth of focus, which can be found by various methods. (*See* for example, Berlage, 1932, p. 504-516). For the first layer, equation (2) does not give d_1, but $d_1 - \frac{1}{2}h$:

$$2d_1 - h = \frac{v_2\tau_2}{\sqrt{\left(\frac{v_2}{v_1}\right)^2 - 1}} = \frac{\tau_2}{\sqrt{\left(\frac{1}{v_1}\right)^2 - \left(\frac{1}{v_2}\right)^2}} = \Delta^* \sqrt{\frac{v_2 - v_1}{v_2 + v_1}} \tag{3}$$

$\Delta^* =$ distance at which a and b (Fig. 1a) intersect. For the second layer, a combination of equations (2) and (3) leads to

$$d_2 = \frac{v_3\tau_3 - (2d_1 - h)\sqrt{\left(\frac{v_3}{v_1}\right)^2 - 1}}{2\sqrt{\left(\frac{v_3}{v_2}\right)^2 - 1}} = \frac{v_3\tau_3 - v_2\tau_2\sqrt{\frac{v_3^2 - v_1^2}{v_2^2 - v_1^2}}}{2\sqrt{\left(\frac{v_3}{v_2}\right)^2 - 1}} \tag{4}$$

From equation (3) it follows that the resulting value of d_1 depends on the assumed value for the depth of focus. h cannot be larger than d_1, as otherwise no waves with velocity v_1 could be observed. Frequently it is found that $h = d_1$ approximately; this is not surprising, as at the discontinuity between two layers the breaking strength is likely to be smaller than in the interior of either layer. On the other hand, the assumed value of h does not affect the calculation of the thicknesses of

the deeper layers; d_2, d_3, and so forth, are usually found in successive steps using equation (2).

There are various sources of error in determining d_1, d_2, d_3, and so forth. An incorrect assumption of h affects d_1 only; if h is assumed too small by an amount x, d_1 is found too small by an amount $x/2$. The intercept times τ are another source of error in finding d_1, d_2, and so forth. Using seismograms of the same shock, different authors or the same author at different times have considered different values for τ as most likely and, consequently, have arrived at different results for the thicknesses of the layers. Especially, if $d_1 = h$, the resulting value of d_1 is proportional to τ_2. As τ_2 is frequently between 1 and 2 seconds, an error of ½ second in the assumed value of τ_2 may change the resulting value of d_1 by about 30 per cent. Therefore, the intercept times τ must be determined as accurately as possible either by plotting the straight lines a, b, c, and so forth (Fig. 1a), or by calculating their parameters, or by plotting the time intervals measured directly on the seismogram. This last method has the advantage that the time corrections do not enter, while the first two can make use of all seismograms with accurate timing in which only one of the impulses is clearly recorded. In Figure 2, time intervals between phases are plotted. The scattering of the points may be due to incorrect identification of the beginning of a phase, uneven speed of the recording drum, or to different structure beneath the stations, including the effect of sediments.

The absolute values of the velocities have less influence on the results than the differences between the velocities. Whether this difference changes locally can be found from plotting the time intervals $t_x - t_y$ between two phases, x and y:

$$x - t_y = (\tau_x - \tau_y) + \left(\frac{1}{v_x} - \frac{1}{v_y}\right)\Delta \tag{5}$$

The distance Δ' at which the two travel time curves intersect is found from

$$\Delta' = \frac{\tau_y - \tau_x}{\frac{1}{v_x} - \frac{1}{v_y}} \tag{6}$$

If, especially, the phase x is the wave through the uppermost layer, then $\tau_x = 0$, and

$$\tau_y = \Delta'\left(\frac{1}{v_x} - \frac{1}{v_y}\right) \tag{7}$$

Equation (7) furnishes an approximate value for τ_y where a few observations give the distance Δ' and where the velocities are known from other sources.

The use of equation (2) leads to correct values for successive thicknesses only, if all layers have been identified. Frequently seismograms indicate a phase between b and c, but the data are insufficient to find v and τ for this phase; or such a phase may exist without furnishing a noticeable impulse on the seismogram, or it may have been misinterpreted as an impulse from a reflected wave or a wave which has travelled partly or wholly as a transverse wave. If a layer with a thickness d and the velocity v_x is inadvertently omitted, and it is assumed incorrectly that the waves travel through that layer with the lower speed v_y of the next shallower layer, the combined thickness of these layers is found too small by an amount given by

$$E = d\,\frac{\sqrt{\left(\frac{v_n}{v_{x-1}}\right)^2 - 1} - \sqrt{\left(\frac{v_n}{v_x}\right)^2 - 1}}{\sqrt{\left(\frac{v_n}{v_{n-1}}\right)^2 - 1}} \tag{8}$$

If, for example, the velocity in the second layer is 6 km/sec, in the third 7 km/sec, and in the deepest, fourth, layer 8 km/sec, and if the third layer is neglected, assuming that the velocity is 6 km/sec down to the deepest layer, the resulting error is about 0.6 d; or 6 km, if $d = 10$ km.

Thus far it has been assumed that the velocity increases with increasing depth. If a layer with lower velocity underlies a layer with higher velocity, the rays arriving at the common boundary are refracted toward the vertical, and all rays arrive at the next deeper boundary. No rays have their deepest point in this layer, and no phase in the seismograms reveals directly the existence of the layer with the low velocity. Waves are reflected from the lower boundary of such a layer, but in seismograms of earthquakes such reflections are rarely identified, even where expected. This fact should not be confused with the conditions in seismic reflection shooting where the methods are especially designed to find such reflections. The error introduced in the calculation of the thicknesses of the layers from the low velocity layer downward is again given by equation (8). It is now negative; all such values of d are found too great.

In regions with the same structure, waves b and c in Figure 1b arrive earlier, when the depth of focus is larger, while the direct wave is mainly affected at short distances from the epicenter. The change δ of the intercept time corresponding to a change Z in depth is given by

$$\delta = -Z\sqrt{\frac{1}{v_1^2} - \frac{1}{v_n^2}} \tag{9}$$

which follows from (2). In most instances, the velocity in layer 1 is about 5.6 km/sec for longitudinal waves (transverse waves are not dis-

cussed in the present paper). In the deepest layer considered here the velocity is close to 8 km/sec. For this layer, equation (9) gives $\delta = -0.13Z$. A decrease in focal depth of about 8 km produces an increase of the intercept time by about 1 second. For an intermediate layer with a velocity of 6.3 km/sec, the same decrease in depth would increase τ_2 only by about 0.65 second.

The effect of a change Z in the thickness of a layer m on the intercept time can be found, too, from equation (2):

$$\delta = 2Z\sqrt{\frac{1}{v_m^2} - \frac{1}{v_n^2}} \tag{10}$$

In this equation the factor 2 should be omitted if the change only occurs either at the station where the shock is recorded or in the region of the source:

$$\delta = Z\sqrt{\frac{1}{v_m^2} - \frac{1}{v_n^2}} \tag{11}$$

The following conclusions may be drawn from equations (9) and (10): If the slope of curves a and b and the time τ_2 are the same in two earthquakes, but c arrives earlier in the second, (τ_3 smaller), the boundary between layers 2 and 3, in general, will be shallower in the second (possibly only near the source or only near the station); if, however, all intercept times are shorter in the second shock, this may have either originated at a deeper source, or all layers are thinner. A decision may possibly be reached by calculation of the depth of focus or by recognizing that the focus cannot be below the boundary between layers 1 and 2, if waves with an apparent velocity v_1 are recorded.

RESULTS FROM ARTIFICIAL EXPLOSIONS

The use of artificial explosions for investigation of the layers in the earth's crust has the advantage that the exact location and depth of the source and, frequently, the origin time are known. On the other hand, the waves through the deeper layers arrive rather late, the distance $\triangle^*$ (Fig. 1a) is greater than in earthquakes, and relatively large energies are needed to record impulses b or c over a distance interval great enough to find the velocities in deeper layers. Moreover, the wave velocities in ancient sediments and in various types of igneous and metamorphic rocks near the surface may exceed the value of 5.6 km/sec, which is usually found in the underlying layer a. Thus, in Ordovician sediments at a depth of 1200 meters, velocities near 6.1 km/sec have been observed, and values up to 7 km/sec in other ancient sediments; 6 km/sec or more have also been found in some gneisses, quartzites, dolomites, rock salts, and in many basic and ultrabasic rocks. In all

instances where an extensive layer of rocks with velocities exceeding 5.6 km/sec lies above granite (with a velocity of 5.6 km/sec or less), no travel time curve indicating the granitic layer can be expected, if the source of the waves is in the upper high velocity layer. (*See* section on theory.) Consequently, if a velocity of about 6 km/sec is found for the direct waves in artificial explosions, as in northern Germany, New England, and southern California, this may belong to one of the rocks just mentioned even though no specific rock with such a high velocity has been identified. It is not possible to conclude in such instances that the layer with a velocity of 5.6 km/sec is absent. In blasts in southern California waves traveling near the surface with velocities of about 6 km/sec have been observed in some areas, of about 5½ km/sec, or both, in others (Wood and Richter, 1931; 1933). Especially, about 6 km/sec have been found between Colton and Pasadena in several recent blasts by C. F. Richter (unpublished). In this same area, no waves traveling near the surface with 6 km/sec have been detected in earthquake records (Gutenberg, 1943), but the velocity of $\overline{P}$ was 5.59 km/sec with a probable error not exceeding 0.1 km/sec in many shocks. The epicenter of one earthquake was only a few kilometers from the source of blasting near Colton. The earthquake waves passing through such high velocity surface layers would arrive a few tenths of a second earlier than usual. These and other differences due to local conditions can be detected only by using a large body of highly accurate data.

In the other areas with observations of high velocities near the surface, no earthquake data are available. The best observed shocks in New England, the New Hampshire shocks of December 1940, unfortunately do not contribute to the solution, as the nearest stations were too distant to furnish positive evidence for or against the existence of a layer with a velocity of 5.6 km/sec in that region, and, furthermore, these shocks probably originated in one of the intermediate layers; in this case no travel time curves corresponding to the velocities in the upper layer exists. The lowest velocity found by Leet and Linehan (1942) for longitudinal waves in these shocks was 6.44 km/sec. Farther to the north, Hodgson (1942) obtained the usual velocity of about 5.5 km/sec for longitudinal waves in the uppermost layer in a preliminary investigation of rockbursts in Lake Shore Mines, Kirkeland Lake, Ontario, recorded at Ottawa, Canada.

GENERAL RESULTS FROM EARTHQUAKES

The records of near-by earthquakes in continental regions show a remarkable similarity (Pls. 1, 2). The differences in the recorded periods

depend mainly on the characteristics of the instruments; those with a long free period have a relatively higher magnification for long waves (with periods close to or above their own) than those with a short free period. In Plate 2 the time scale was made approximately the same for all seismograms, but noticeable deviations may exist, as in some of the originals the correct length of 1 minute could not be ascertained. Figure 2 is based on readings made at the stations where the shocks were recorded.

In general, the velocity of longitudinal waves in the first layer below the sediments is about 5.6 km/sec. These waves (corresponding to *a* in Figure 1) are usually indicated by the symbol $\overline{P}$, sometimes by *Pg*, where the "*g*" indicates "granite", the material considered most likely to prevail in this layer.

In some regions, there seem to be several intermediate layers (corresponding to the one, number 2 on Figure 1) with velocities between 6.0 and 7.8 km/sec. The symbols differ with the writer: P^* used by their discoverer (Conrad, 1925); *Px, Py, Pm, Ps; P* with an index giving the velocity, for example, $P_{6.05}$. In the present paper, the author has used the symbol P^* for all of them. For a few regions, time differences $\overline{P}$-P^* are plotted on the right side of Figure 2. The slopes of the curves are definitely different; from equation (5) it follows that the velocities of P^* are different, since those of $\overline{P}$ are nearly the same in these cases.

The boundary between the lowest intermediate layer and the deeper material (probably ultrabasic) is frequently called the "Mohorovičić discontinuity" in honor of its discoverer (Mohorovičić, 1910). Practically everywhere the velocity below this discontinuity is near 8.0 km/sec within the limits of error (Gutenberg, 1927a; 1932; 1943; Jeffreys, 1937; Caloi, 1939a; Macelwane, 1939). This is confirmed by Figure 2, left side, which shows observed time differences between the wave $\overline{P}$ through the uppermost layer and this wave (symbol *Pn*) through the "ultrabasic" layer. For all regions of the earth, for which data are available, these time differences *P-Pn* plotted as a function of epicentral distances are on practically parallel lines. In Figure 2, the line plotted to fit the data for the western part of southern California is repeated in all sections on the left side as a barred line to make a comparison easier. From equation (5) it follows that the difference of the reciprocals of the two velocities involved is constant.

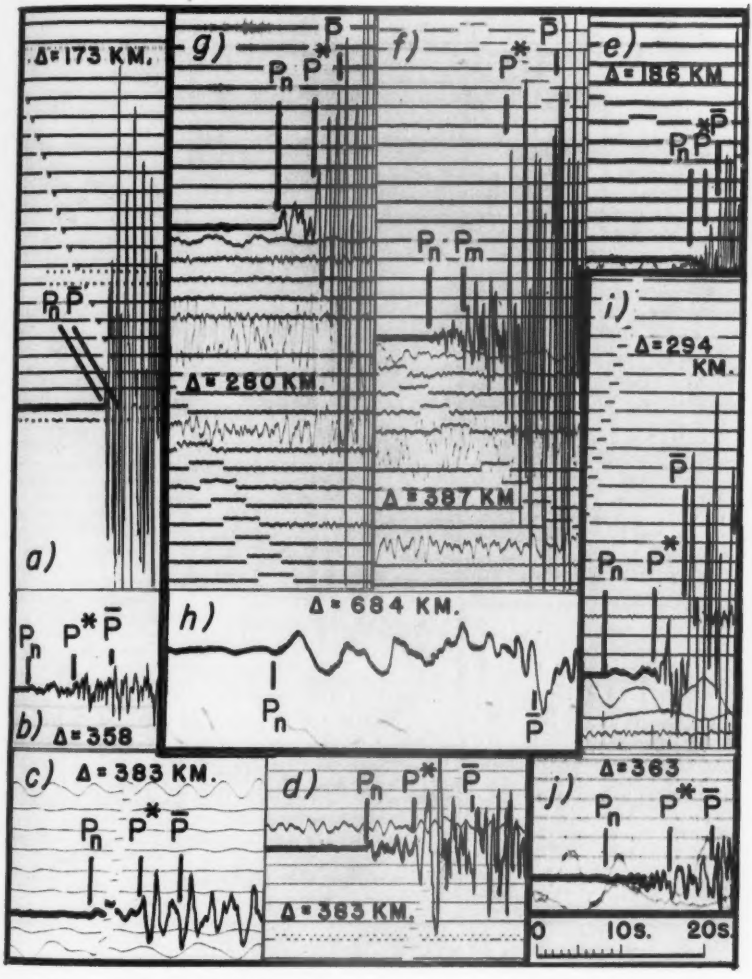
The material immediately below the Mohorovičić discontinuity is probably crystalline. There are indications that the melting point is reached at a depth of roughly 80 km (Daly, 1940, p. 18; Gutenberg and Richter, 1939a).

Only continental areas are considered here. In most of the Pacific, the

TABLE 1.—*Data concerning Plates 1 and 2*

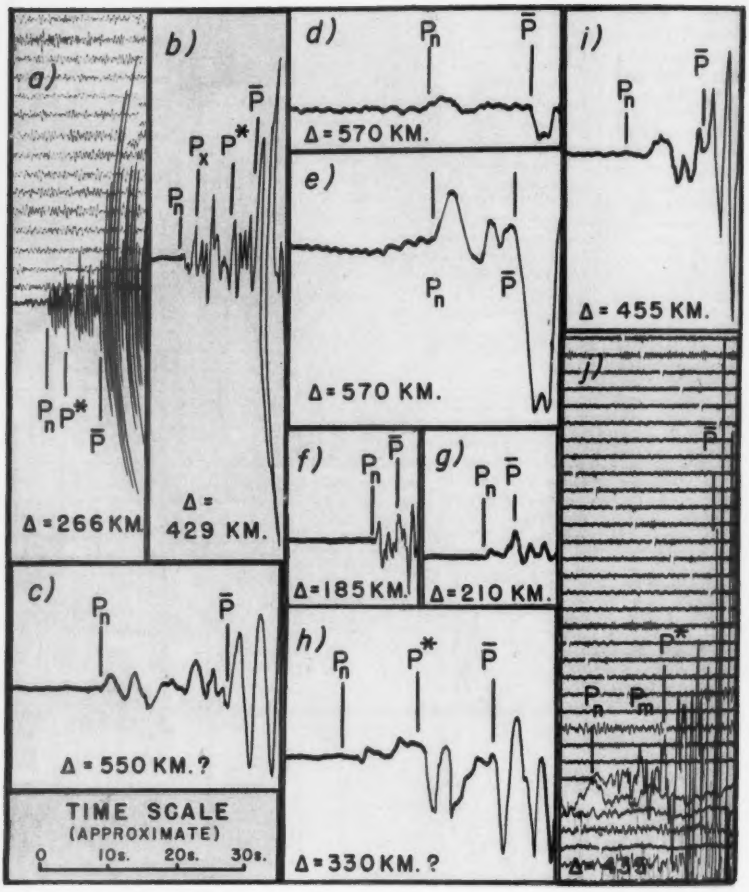
Plate No.	Region of epicenter	Station	Instrument	Taken from
1a	Northern Owens Valley, Calif.	Haiwee, Calif.	Standard torsion E–W (Wood-Anderson)	Original
1b		Santa Barbara, Calif.	Standard torsion N–S	
1c		Pasadena, Calif.	Long period torsion E–W	
1d		Pasadena, Calif.	Standard torsion N–S	
1e		Santa Barbara, Calif.	Standard torsion N–S	
1f	Long Beach, Calif.	Tinemaha, Calif.	Standard torsion E–W	
1g		Haiwee, Calif.	Standard torsion N–S	
1h		Tucson, Ariz.	Long period torsion N–S	Original*
1i	Imperial Valley, Calif.	Pasadena, Calif.	Standard torsion N–S	Original
1j	Imperial Valley, Calif.	Boulder City, Nev.	Standard torsion E–W	Original*
2a	Belgium	Bochum, Germany	Wiechert 17,000 kg. N–S	Somville (1939)
2b	Belluno, S. Alps, Italy	Neuchâtel, Switz.	De Quervain 20,600 kg. E–W	Caloi (1938)
2c	Southern Apennines, Italy	Zagreb, Yugoslav.	Wiechert 1000 kg. NE–SW	Mohorovičić (1910)
2d	Leninakan, south of Caucasus	Baku, USSR	Galitzin E–W	Raĭko (1930)
2e	Leninakan, south of Caucasus main shock, 3 hrs. earlier.	Baku, USSR	Galitzin E–W	Raĭko (1930)
2f	Central Asia near 41° N. 74° E.	Andijan, USSR	Galitzin (mechanic) E–W	Rozova (1936)
2g	Central Asia, near 41° N. 74° E.	Frunse, USSR	Galitzin (mechanic) E–W	Rozova (1936)
2h	Pahiatua, New Zealand	Takaka	Imamura strong mot. E–W	Hayes (1937)
2i	Tazima, Japan	Tokyo	Wiechert 200 kg. E–W	Wadati (1926)
2j	Imperial Valley, Calif.	Santa Barbara, Calif.	Benioff short period vertical	Original

* Record of Tucson by courtesy of the U. S. Coast and Geodetic Survey, record of Boulder City by courtesy of the Bureau of Reclamation, National Park Service, and of the U. S. Coast and Geodetic Survey.



CHARACTERISTIC SEISMOGRAMS (BEGINNING) OF NEAR-BY SHOCKS IN CALIFORNIA

For details, *see* Table 1. Some of the figures show, besides, waves from after shocks or other earthquakes, (a–d) source in northern Owens Valley; (e–h) near Long Beach; (i, j) in Imperial Valley.



SEISMOGRAMS OF NEAR-BY SHOCKS IN VARIOUS REGIONS
For details, *see* Table 1.

granitic and intermediate layers seem to be practically absent; in the Atlantic and Indian Oceans, they seem to be noticeably thinner. (For details, *see* Gutenberg and Richter, 1939b.)

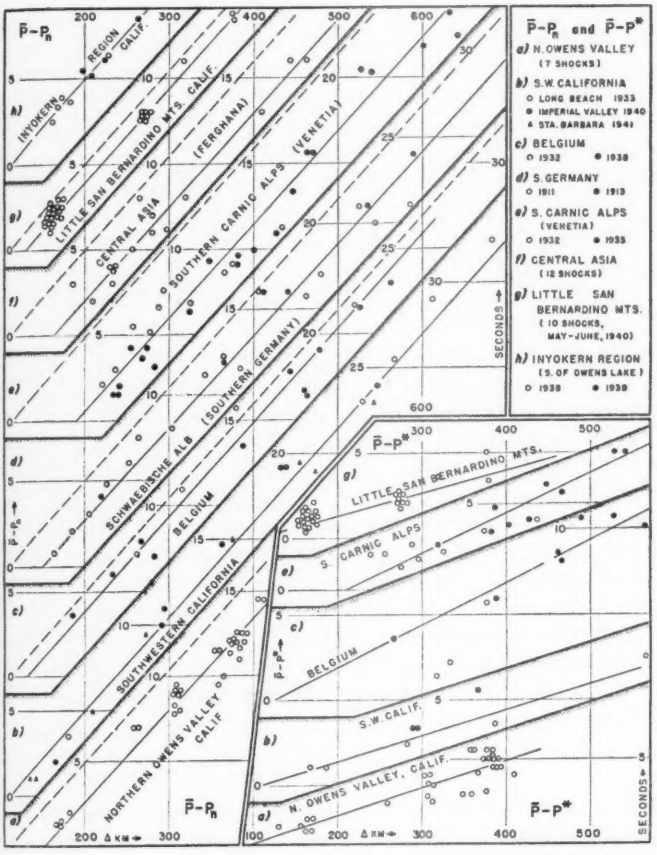


FIGURE 2.—*Travel time differences between various longitudinal phases*

Left side, $\bar{P} - Pn$; right side, $\bar{P} - P^*$

Whereas practically all seismologists agree that within the uppermost and in each intermediate layer the velocity is nearly constant and probably increases slightly with depth, Schmerwitz (1938; 1939; 1940) concluded that the velocity decreases gradually from about 5.9 km/sec near

the surface to about 5.5 km/sec at the Mohorovičić discontinuity at a depth of about 50 km. He based his findings on a combination of the velocity of about 6 km/sec found in blasts (discussed in the preceding section) and results from earthquakes. For some shocks he calculated the co-ordinates, the depth of focus, the origin time, and the velocity of the waves in one step by using the method of least squares. As in the shocks under consideration the travel times, at distances greater than about 5 times the depth of focus, are affected very little by the focal depth, and data from shorter distances are usually scarce, small errors may affect rather strongly the resulting depth of focus (Jeffreys, 1939). Too great a depth of focus usually leads to too early an origin time. If H in Figure 1b is taken deeper below the surface, the corresponding time HE is increased, 0 is taken earlier, and the slope of line a in Figure 1a is increased. The greater this slope, the smaller the velocity which results for layer 1. Consequently, if the depth of focus is found greater than it really is, the velocity of the waves in the upper layer frequently is found too small. Thus the results of Schmerwitz, that waves from a deeper focus have a lower velocity, may easily be explained as a spurious result of the interrelation of the errors in the determination of the two quantities. For this reason, the depth of focus is better determined separately, for example from records at near-by stations only, as is usually done. No indication of a gradual decrease in velocity with depth in the upper layers has been found by this method. If Schmerwitz were right, $\overline{P}$ could be observed only at rather short distances, depending on the depth of focus h; in particular all rays from sources at the surface must then be curved downward until they reach the Mohorovičić discontinuity, and only diffracted $\overline{P}$-waves could be observed on blast records, aside from Pn which would arrive after many seconds. Actually, the $\overline{P}$-waves from blasts are too strong to be explained as diffracted waves.

DATA FOR EUROPE OUTSIDE ALPS

Records of two earthquakes in England (Jeffreys, 1927; 1933) and two in the English Channel (Mourant, 1931) gave velocities for $\overline{P}$ near 5.5 km/sec, for P^* near 6.4, and for Pn between 7.6 and 8.2 km/sec. However, the values for the intercept times τ are uncertain by several seconds in most instances, and no approximate thickness can be found for any layer. The records of two earthquakes in Belgium (Gees, 1937; Somville, 1939) show about equal travel time differences for a given epicentral distance (Fig. 2c). In the smaller shock, of 1932, apparently two impulses given by Gees as P^*, belong to a different phase. For the

1932 shock, the depth h has been found independently by three authors between 23 and 35 km; for the 1938 shock, Schmerwitz (1940) found $h = 67 \pm 15$ km. From Figure 2c, we find approximately

$$\overline{P} - P^* = 0.024\Delta - 2\tfrac{3}{4} \qquad \overline{P} - Pn, = 0.053\Delta - 5\tfrac{1}{2}$$

The velocities of 5.6, 6.4, and 8.0 km/sec respectively for the three layers correspond to these differences as well as to the direct findings by other authors (5.48; 5.49; 5.6; 5.62; 5.7 for $\overline{P}$; 6.42 for P^*; 7.63; 7.97; 8.0 for Pn) for the two shocks. Equation (3) gives $(2d - h) = 30$ km (excluding $h = 67$ km). Apparently the depth of focus was close to the lower boundary of the first layer, which is common, and approximately $d_1 = h = 30$ km. Equation (4) gives $d_2 = 10$ km and the depth of the Mohorovičić layer in Belgium at about 40 km. Similar results are found from records of an earthquake originating west of Köln (Cologne). The data are rather scanty but were sufficient to indicate a smaller depth of the Mohorovičić discontinuity there than in the following group (Gutenberg, 1929).

The largest shocks in Germany in recent decades have occurred in the Schwaebische Alb (north of the upper Danube). Data for two shocks, 1911 and 1913, have been published with seismograms by Gutenberg (1915). Only a few clear P^*-phases are available; they give approximately the same values for the first layer as those in the Belgian shocks. However, the Pn-waves arrive decidedly later, indicating a greater thickness of the intermediate layer or layers. From Figure 2d, we find $\overline{P} - Pn = 0.055\Delta - 7\tfrac{3}{4}$; the constant term, which is most important, has been found by Jeffreys (1933) to 8.3 seconds for the larger and 6.3 seconds for the smaller shock. All velocities are about the same as in the preceding cases. The thickness of the first layer is probably between 25 and 30 km, that of the second between 20 and 25 km with a total above rather than below 50 km. Other shocks originating in this region have been investigated by Hiller (1935; 1936), who found about 20 to 30 km for the thickness of the first layer and about 45 for the depth of the Mohorovičić discontinuity.

Yugoslavia is the only other region of Europe outside the Alps for which data are available. Mainly from a shock in the Kulpatal, Mohorovičić (1910) found velocities of 5.6 for $\overline{P}$ and of 7.8 km/sec for Pn, beginning at a depth of about 50 km. For a shock in 1923, near Imotski, Tillotson (1931) included data for the intermediate layer; according to his results, the velocities are 5.5, 6.3, and 7.8 km/sec respectively, that is, about the same as in northwestern Europe, and below a sedimentary layer, 4 km thick, the layers have thicknesses of 13 and 25 km respectively, giving a total of 42± km. The present author has found

approximately the same total amount by plotting time differences for both shocks, which gave $\overline{P} - Pn = 0.055\Delta - 6$ sec.

THICKNESS OF LAYERS UNDER ALPS AND THEIR FORELAND

In Figure 2e the differences $\overline{P} - Pn$ and $\overline{P} - P^*$ are plotted for two shocks in the southern Carnic Alps (Caloi, 1932, 1938); Plate 2b gives a seismogram for waves traveling through and under the Alps. The curves in Figure 2e indicate that P^* as well as Pn were noticeably later in these shocks than in any other instance anywhere for which data are available. A depth of focus of about 20 km was found by Caloi (1939b), and we must conclude that the late arrivals of Pn and P^* are due to a greater thickness of the layers, especially the first. The difference involved is still greater than it seems at a first glance, as many stations used in this and the previous instances are the same, so that most of the difference must be accounted for by the conditions near the source. Equation (11) gives a thickness of roughly 40 km for the first layer and a depth of about 60 km for the Mohorovičić discontinuity. The velocities found by Caloi are about 5.7, 6.5, and 7.9 km/sec and agree well with the differences found from Figure 2e: $\overline{P} - P^* = 0.024\Delta - 5$ sec and $\overline{P} - Pn = 0.055\Delta - 10$ sec. His resulting depths are slightly smaller than those just mentioned due to his assumption that they are equal at the source and at the stations.

Other shocks from the Alps or their immediate foreland include one originating near Schwadorf (near Vienna), one in the Tauern, one in northern Tyrol, and one in the Vispertal. The first two have been studied by Conrad (1925; 1928), the third by Graefe (1932), and the last by Wanner (1930). Some have been rediscussed, for example by Jeffreys (1933), Caloi (1939), and Schmerwitz (1938; 1939). Nine determinations of the focal depth h by six different authors, including all shocks, but most for the third, agree within the limits of errors and give values between 26 and 38 km. Other data for these shocks (in parentheses, no. of results, no. of authors) are: v_1 between 5.4 and 5.72 (8; 4); v_2 between 6.25 and 6.7 (4; 3); v_3 between 7.65 and 8.3 (9; 5); τ_3 between 6.1 and 9½ (7; 4). By plotting the time differences, the present author finds for the Schwadorf shock: $\overline{P} - P^* = 0.023\Delta - 3$ sec, $\overline{P} - Pn = 0.054\Delta - 9$ sec; for the Tauern shock: $\overline{P} - Pn = 0.052\Delta - 7½$ sec; for the Tyrol shock: $\overline{P} - Pn = 0.053\Delta - 8$ sec; the P^* observations for these two shocks are too scanty for separate use, and this is true, too, for all differences in the Vispertal shock. The differences for the shocks are within the limits of error, and all data combined give about 35 km for the thickness of the first layer (Conrad found 40 km),

and about 55 to 60 km for the depth of the Mohorovičić discontinuity. (Conrad gave about 50 km, supposing the same value under the source and the stations.)

A summary of the main results for Europe is given in Table 2. While the details are only approximate, and the results are samples rather

TABLE 2.—*Approximate thickness d_1 of the upper ("granitic") layer, d_2 of intermediate layers and depth D of Mohorovičić discontinuity in Europe*

Region	Thickness in km (approx.)		
	d_1	d_2	Total (D)
Northwestern Europe	30	10	40
Schwaebische Alb	25–30	20–25	50
Northern Alps, and Vorland	35	20–25	55–60
Southern Carnic Alps	40	20–25	60–65
Yugoslavia	15	25	40

than averages for the regions mentioned, for Europe the maximum depth of the Mohorovičić layer is undoubtedly under the region of the Alps, which means that the Alps have a root. This root is mainly a result of a greater thickness of the uppermost (granitic) layer. More and better data on the velocities in the intermediate layers are necessary to find out how far an increase in the thickness of the intermediate layers contributes to the result.

There are probably real differences in the velocity of P^* in various parts of Europe (Jeffreys, 1933; 1937), and there seem to be additional phases with higher velocities (for example, about 7 km/sec (Stoneley, 1931); or about 7.8 km/sec, Conrad, 1928, or Jeffreys, 1933). Equation (8) shows that the thickness of the intermediate layers could be greater by about 5 km or even more in such a case.

CRUSTAL LAYERS IN ASIA

While no data are available for the thicknesses of the layers and for the velocities in South America and Africa, some information exists for Asia. Raïko (1930) found a velocity of about 5¾ km/sec in the uppermost layer for waves traveling mainly through the foreland of the Caucasus; and the few data and reproduced seismograms (Pl. 2d, e) with well-recorded $\overline{P}$ and Pn phases through that region, for example, several with $\overline{P} - Pn = 8$ seconds at a distance of about 300 km, suggest

the possibility of a depth of the Mohorovičić discontinuity of roughly 50 km, similar to the foreland of the Alps.

The layering in the Ferghana region, central Asia, has been investigated by Rozova (1936). Errors (Rozova, 1940) are introduced not only because of the rather small number (six) of stations but also because of the occurrence of shocks at various depths (range almost 300 km). Data for 12 carefully selected shocks are plotted in Figure 2f; Plate 2f, g, gives examples of seismograms. Rozova's findings ($v_1 = 5.54$ km/sec, $d_1 = 20$–25 km; $v_2 = 6$ km/sec; $D = 45$ km; $\tau_3 = 9$ sec; $v_3 = 7.82$ km/sec) and Figure 2f indicate that the structure there is similar in all respects to that in the foreland of the Alps.

Although there are many seismological stations in Japan, the data available for our purpose are rather scanty. In the earlier years, the time corrections were well known at only a few stations, and in recent years few data have been published. Combined findings by Gutenberg (1925), Suda (1925), Wadati (1926), Matuzawa (1928; 1929), Matuzawa *et al.* (1929), Hodgson (1932), and Imamura (1937, p. 285) indicate that the structure in Japan corresponds roughly to the structure in southern Germany. For the Tazima earthquake in 1925 (seismogram at Tokyo in Pl. 2i) and the Sagami Bay shock in 1923, the velocity in the upper layer has been found to be near 5.6 km/sec, for other shocks between 5 and 6.3; however, in these shocks even such differences may be due to errors. The velocity of P^* seems to be near 6.2 km/sec. For *Pn*, values between 7.5 and 7.8 km/sec have been given, but a value near the upper limit is more probable, as most instruments in Japan are adjusted to low sensitivity so that they may record the rather frequent large shocks, and in consequence in many records the small beginning of *Pn* is certainly not visible. The travel time curves for $\overline{P}$ and *Pn* intersect in all well-recorded shallow shocks at distances between 100 and 150 km. τ_3 is probably near 7 seconds, and a total thickness of about 45 km of the layers would agree very well with all data. Possibly the structure may differ noticeably in various parts of Japan; Suda (1925) thought that the layers are thinner on the east side than near the Japan Sea. However, this and other supposed differences may be influenced by incorrect or uncertain data.

STRUCTURE IN NEW ZEALAND

Data on wave velocities and structure in northern New Zealand (Pl. 2h) have been published by Hayes (1936) and by Bullen (1936; 1938; 1939). They leave no doubt that the depth of the Mohorovičić layer in New Zealand is smaller than in any other continental region for which we have data. *Pn* precedes $\overline{P}$ from a distance of about 70 km on. Bullen

(1939) finds two upper layers with velocities of 5.0 and 5.55 km/sec respectively and a total thickness of about 10 km, one or two intermediate layers with velocities near 6¼ km/sec, extending down to the Mohorovičić discontinuity at a depth of about 30 km. Below, the velocity is about 8.1 km/sec, as elsewhere.

CRUSTAL LAYERS IN CALIFORNIA

The relatively large number of stations in California (about 15) with very sensitive instruments, high and well-controlled drum speed (1 mm/sec), and frequently recorded radio time signals (Pl. 1a, d) are very favorable for the investigations of structure. Contrary to a statement by Schmerwitz (1938, p. 387), the findings are at least as good as those for Europe, where at best eight stations have equally good equipment. At most of the European stations the magnification of the instruments for short periods and the drum speed are smaller, and this together with irregularities in the drum speed at some stations yield less accurate readings.

For central California, Byerly and Wilson (1935) and Byerly (1938; 1939) found velocities of about 5.6 km/sec for the upper layer, 8.0 below the Mohorovičić discontinuity, and various velocities in the intermediate layers. Jeffreys (1940) investigated carefully the velocity below the Mohorovičić discontinuity in California shocks in general and found it within the limits of error equal to that in Europe. Byerly found an intercept time of *Pn* between 5.3 and 6.3 seconds in all of five shocks in central California for which sufficient data were available. His resulting values, $d_1 = 9$ km, $D = 32$ km, may be slightly too small on account of the larger velocities in some of the intermediate layers and the possibility that the depth of focus was larger than the 3 km which he considered. In any case, his results indicate that the uppermost layer in central California as well as all crustal layers together are thinner than the corresponding layers anywhere in Europe. Further, the intermediate layers, at least in part, consist of different materials.

Gutenberg (1932) had reached similar conclusions for southern California. According to these earlier results, the uppermost layer is about 14 km thick ($v = 5.55$ km/sec) on the average, followed by intermediate layers with velocities of 6.05, 6.8, and 7.6 km/sec below which, at $D = 39$ km, the usual velocity of about 8 km/sec begins. Recent investigations (Gutenberg, 1943) based on 50 shocks in southern California, 1933-1941, confirm these results. In most of southern California, especially in the southwestern part, the thickness of the uppermost layer ($v = 5.58$ km/sec) is found near 18 km, followed by intermediate layers, about 22 km thick (velocities mainly 6.05, in some areas also about 7

and 7¾ km/sec). Below their boundary at about 40 km, the velocity is about 8.00 km/sec. Figure 2b shows (left) differences $\overline{P} - Pn$ and (right) $\overline{P} - P^*$ for shocks in the coastal area and in Imperial Valley; Figure 2h shows the same for two shocks south of Owens Lake. For these two figures, the differences $\overline{P} - Pn$ agree for any given distance; the values of $\overline{P} - Pn$ are the largest of all plotted in Figure 2, indicating relatively early arrival of *Pn* and a relatively shallow Mohorovičić discontinuity. On the other hand, the values of $\overline{P} - P^*$ plotted against distance show a slope different from that in European shocks and indicate a different material in the layer corresponding to P^*. This wave, therefore, has been denoted previously by a different symbol, *Py*, for southern California. There can be little doubt about the difference in the velocities of P^* in Europe and of this wave in California, as it is usually a well-defined phase (Pl. 1). No differences in the lines for $\overline{P} - P^*$ beyond the limits of error have been found thus far in any part of southern California, including the Little San Bernardino Mountains and the Sierra Nevada region (Fig. 2a 2g, right; seismograms, Pl. 1a-d). This means that probably in all these areas the depth of focus and the thickness of the layer immediately below the sediments are approximately the same. There are three shocks in which the nearest station is less than 10 km from the epicenter, and three where this distance is between 10 and 23 km (Gutenberg, 1943, Table 4); in all these instances the resulting depth of focus is near 18 km (extreme values 13 and 24 km). From this fact and from the observed values of $\overline{P} - P^*$ we may conclude that in most of southern California the thickness of the layer in question is about 18 km and that most of the shocks originated near its lower boundary.

On the other hand, Figures 1a and 1g leave no doubt that the *Pn* waves arrive about 1 second later than usual from origins beneath the Little San Bernardino Mountains, and at least 2 seconds later from origins beneath northern Owens Valley (Long Valley). From equation (11) it follows that in the region of the first group of epicenters (near 34° N., 116.3° W.) the Mohorovičić discontinuity is about 10 km deeper than is usual in southwestern California, that is, $D = 50\pm$ km, and that in northern Owens Valley (near 37.6° N., 118.7° W.) it is even 20 to 25 km deeper, $D = 60$ to 65 km. Again, we find an effect which indicates mountain roots.

No shocks are available with an epicenter in the Sierra Nevada proper, and records are insufficient to locate them accurately enough for our purpose. However, Byerly (1925) has found that waves from shocks in the Coast Range region arrive with a delay at the Owens Valley

DELAY OF Pn WAVES IN SIERRA NEVADA REGION
Base map by courtesy of Department of Geology, Stanford University.

stations and "that there exists some barrier to these waves." Later, Byerly (1938; 1939) concluded from additional data that the delay is due to the root of the Sierra Nevada and that this root "projects into the mantle beneath the surface layers by an amount between 6 and 41 km." The conditions are sketched in Figure 1c. At *L* the waves arrive at the normal time; at *M* and *N* they arrive with a delay. The change in thickness given by equation (11) corresponds approximately to the point *S*. Beyond *SN* is a shadow zone; the point *P* does not receive any energy from waves *c* through layer 3. If it is assumed that the first wave arriving at *P* is of the *c*-type, too great a delay may be assumed, and too deep a root may be found.

In Plate 3 the delays found by the author (Gutenberg, 1943) are indicated for each path which was of interest, and data from shocks in the Coast Ranges found by Byerly are added. The rays in the upper right of the figure radiate from the region of the earthquake epicenters in northern Owens Valley. The two centers of rays farther to the southeast correspond to the stations at Tinemaha and Haiwee. In the first case, the rays go from the source to the various stations; in the second and third, they originate at various epicenters. Owens Valley passes through the three centers. The well-expressed valley branching from Owens Valley and looking like an extension of central Owens Valley towards the north-northwest bears no general name on maps; it is sometimes called "Benton Valley". To the east of it are the White Mountains.

From Plate 3 it is evident that the root of the Sierra Nevada does not continue with an appreciable thickness across central Owens Valley, but from Figure 2a we know that it includes northern Owens Valley. Benton Valley could be the approximate boundary of the root; however, the data indicated by the two heavy lines in the upper right corner of Plate 3 are not sufficient to show how far this root extends beyond northern Owens Valley to the east, especially whether it continues under the White Mountains, since the delay may be produced on a short section of the wave path. On the other hand, barred lines in a region not too far from an epicenter or a station exclude an appreciable root there. The root seems to be developed best in the central and possibly the northern Sierra Nevada. In Figure 3 a tentative section across the Sierra Nevada is given.

The root decreases in the southern Sierra, and no indication is found farther to the southwest under the Tehachapi Mountains. Only a few of many available lines have been drawn across the Coast Ranges. All data for this region confirm Byerly's findings that no delay exists either

under the Coast Ranges or under the San Joaquin Valley. On the other hand, the upper (granitic) layer seems to have approximately the same thickness (18 km) under the Sierra as in all other parts in southern California; its local variations cannot be found as accurately as differences in the depth of the Mohorovičić discontinuity (Gutenberg, 1943).

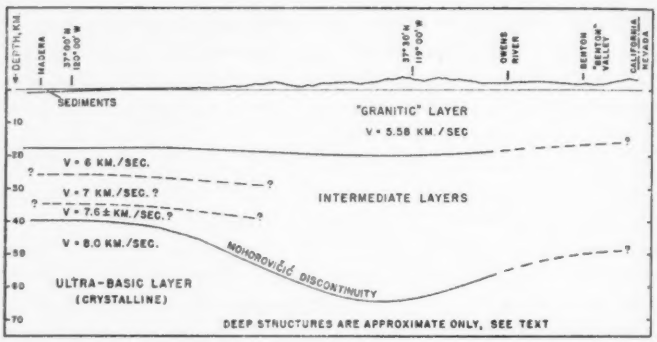


FIGURE 3.—*Tentative section across the Sierra Nevada (true scale)*

The results of a more detailed study, now in progress, seem to exclude a root of the granitic layer extending below a depth of 30 km under the Sierra.

The travel times of other phases traveling under the Sierra Nevada have not yet been investigated in detail. All results available indicate that the root of the Sierra Nevada is due rather to an increase in the thickness of the deeper intermediate layers than in the thickness of the uppermost (granitic) layer. In this respect, the difference between the Sierra Nevada and the Alps is marked.

No adequate data are available for other parts of North America. The fact that at St. Louis, Little Rock, and Chicago the first longitudinal waves from certain distant shocks have arrived as much as 4 seconds early (Lee, 1937; Gutenberg and Richter, 1938) suggests unusual structure there of the deeper crustal layers and unusually high average wave velocities.

For New England, Leet (1936; 1938; 1941) has combined data from blasts with those from a few earthquakes. Leet (1941) found tentatively a depth of 36 km for the Mohorovičić discontinuity and a velocity of 8.4 km/sec below it. As pointed out in discussing results from surface sources, the use of such records will frequently lead to misinterpretations where no earthquake records from short epicentral distances are avail-

able. Therefore, more homogeneous data from New England are needed to obtain more reliably the velocities in the various layers and their thicknesses in New England.

Waves from rock bursts in the Lake Shore Mines, Kirkland, Ontario, recorded at Ottawa, Canada, did not show such peculiarities. In a preliminary investigation, Hodgson (1942) obtained a velocity of 8.1 km/sec below the Mohorovičić layer and suggested a structure similar to that in the average region.

MOUNTAIN ROOTS AND ISOSTASY

There are two fundamentally different basic hypotheses concerning the mechanism of isostasy, one due to Airy and the other to Pratt. (*See* Heiskanen, 1936; Daly, 1940.) The first assumes that the crust of the earth has approximately the same density everywhere, that it "floats" on the substratum, and that thus the visible mass of mountains is compensated by roots, extending into the heavier substratum. The second supposes that the density of the crustal layer is the smaller the higher its surface and that above a certain depth, everywhere the same, the "level of compensation", columns of a given diameter have approximately the same weight everywhere. Differences in the elevation of large units of the earth's crust are due to different expansion or contraction in thermal or chemical processes.

Airy's and Pratt's hypotheses, as stated above, deal with ideal cases. In the actual earth, there are conditions consistent with both, acting either simultaneously or in sequence; however, the mechanism maintaining isostasy seems to be mainly of the Airy type (Gutenberg, 1927b). The Pratt hypothesis assumes processes which, operating chiefly during revolutionary periods in the earth's history, have changed the height and density of a given column of rock regardless of the existence or nonexistence of equilibrium and without affecting the existing isostatic conditions; his hypothesis tends to explain observed facts rather than processes now in action with observable effects. The Airy hypothesis assumes plastic flow in regions where the equilibrium is disturbed; it takes thousands of years to approximately restore the equilibrium after the disturbing process has ceased. The main changes in the process during the history of the earth are connected with the cooling of the earth, producing an increase in strength and a decrease in plasticity.

The findings of this paper are in agreement with the basic idea of Airy's hypothesis. However, important details of this hypothesis are only rough approximations. The compensation involves roots of several layers with regionally different thickness, wave velocity, elastic con-

stants and density. From the literature it seems that the average density of the earth's crust in mountain regions frequently is smaller than in lowlands; the surface layer has the greatest density in oceanic areas in agreement with Pratt's hypothesis. A detailed investigation of this problem is needed, since published results are scanty and in part doubtful. The Mohorovičić discontinuity, which forms the boundary of the mountain roots discussed here, is not identical with the surface of the asthenosphere. The material between these two discontinuities is considered crystalline. Since the roots of the mountains probably influence the isotherms below them, we must expect many irregularities of the surface of the asthenosphere; the information available is inadequate to estimate if and how far such differences in the surface of the asthenosphere contribute to isostatic equilibrium. There is good reason to assume that the strength in the vitreous asthenosphere is smaller than at the Mohorovičić discontinuity, but the difference in density is unknown; a smaller density below the surface of the asthenosphere than above it would require a "reversed root" there to compensate too large a mass above.

SUMMARY

All results based on the study of seismograms of near-by continental earthquakes indicate that below the sediments is a layer with a velocity in general between 5.55 and 5.60 km/sec for logitudinal waves. It is considered granitic. Below it intermediate layers with higher velocities definitely show regional differences. Their lower boundary is the "Mohorovičić discontinuity", below which a material, probably ultrabasic, with a velocity of about 8.0 km/sec is found nearly everywhere. The velocities through the granitic as well as those through the ultrabasic layers are the same in different regions, within ± 0.4 km/sec.

The maximum thickness (60 to 70 km) of the "continental layers" above the Mohorovičić discontinuity thus far has been found under the southern Alps and about the same under the Sierra Nevada. Whereas the root of the Alps seems to be due mainly to an increase in the thickness of the granitic layer, present indications are that the granitic layer under the Sierra Nevada extends roughly down to the same depth (about 20 km) as generally in southern California; mainly the intermediate layers are thicker under the Sierra. In other continental regions, the Mohorovičić discontinuity is in general at a depth of about 50 km under areas with moderately high mountains, and of about 40 km near oceanic coasts. An especially low value of 30 km has been found in New Zealand. Under the Atlantic and Indian Oceans these layers have

probably an even smaller total thickness and are practically absent under the Pacific Basin. The boundary of the crystalline crust is within the ultrabasic material.

Many more data are needed to bring out details in this rough picture of the earth's crust.

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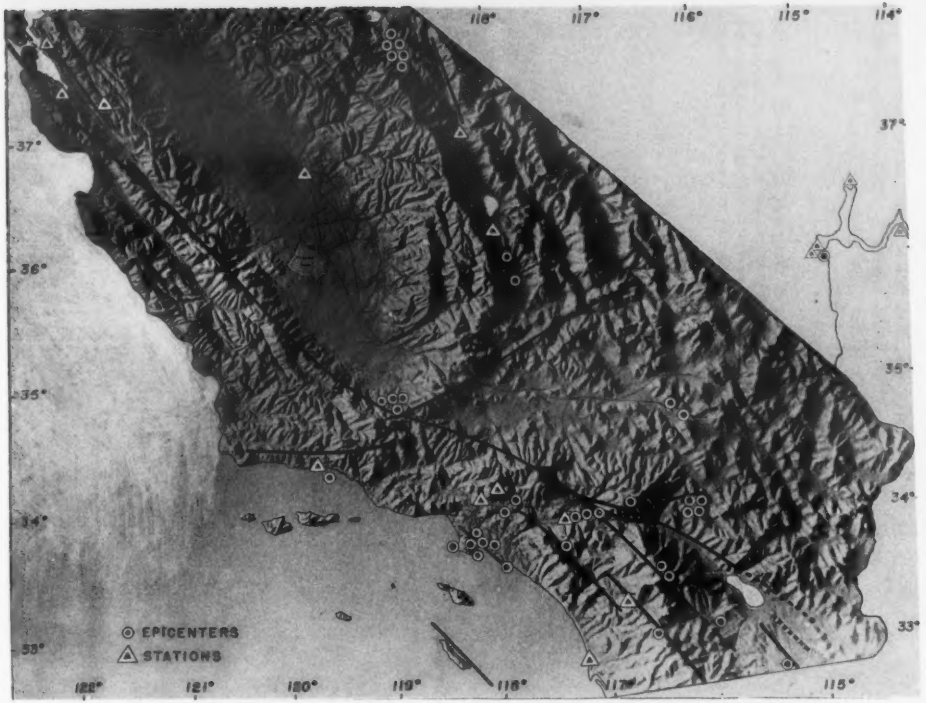

MAP OF STATIONS AND EPICENTERS

Base map by courtesy of Department of Geology, Stanford University. Fault traces after Willis (1923) and Jenkins (1938).

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EARTHQUAKES AND STRUCTURE IN SOUTHERN CALIFORNIA

BY BENO GUTENBERG

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ABSTRACT

During recent years the accuracy in determining the arrival times of earthquake waves in southern and central California has been notably improved, so that more detailed results concerning the structure of the earth's crust in California are to be expected from the study of such records. Fifty shocks were selected, and their epicenters, depths of foci, origin times, and magnitudes determined and discussed. All known corrections have been applied; in particular, the effect of the altitude of some of the stations is beyond the probable error of the readings. With this improved material it is possible not only to investigate more accurately than heretofore the wave velocities in the various layers, but also differences in structure, especially local effects near the stations.

INTRODUCTION

About 10 years ago the author published data concerning 21 earthquakes in southern California[1] (Gutenberg, 1932a). In the interim, additional seismograms of shocks suitable for a detailed study have accumulated. The new material is much superior to that used previously for several reasons. About 1934 the installation of highly sensitive Benioff vertical seismographs at all stations of the Pasadena group was completed; in 1938-1939 the stations were equipped with new chronometers and with apparatus to record automatically radio time signals 6 times a day. In 1939, a station at Palomar was added.

Very valuable additional information came from the stations in central California (central station Berkeley), for which Doctor P. Byerly kindly furnished many seismograms. The main improvements for our purpose were the installation of a short period vertical Benioff seismograph at Berkeley and the installation of a Wood-Anderson torsion seismograph at Fresno. The comparison of seismograms recorded at Fresno, to the west of the Sierra Nevada, and at Haiwee and Tinemaha to the east of the Sierra, makes it possible to determine more accurately the effect of the Sierra on the arrival times; usually, the path to at least one of the three stations does not pass under the Sierra, and the beginning on such records can be used to determine the epicenter of the shock without the risk of using a wave delayed by the root of the Sierra.

The four stations near Lake Mead, operated by the Bureau of Reclamation, National Park Service, and the U. S. Coast and Geodetic Survey, furnished valuable data. Records were kindly lent by Doctor Carder. Short period Benioff instruments have just been installed at three of these stations. Their future records will doubtless not only clear up questions of a local nature but will also permit a more detailed study of shocks in the region of the Sierra Nevada, southwestern Nevada, and eastern California.

Finally, a short period and a long period Benioff vertical seismograph were installed at Tucson in 1936. Records of these and the torsion instruments there were kindly lent by the U. S. Coast and Geodetic Survey. Fortunately, the ground at that station is very quiet, so that a high magnification for the short period Benioff instrument can be used. In the writer's judgment, this instrument is operated with the highest magnification used anywhere in continuous recording of earthquakes—between one quarter and one half million for waves with periods of a few tenths of a second. The *Pn*-waves of moderate shocks many hun-

[1] By agreement with Dr. P. Byerly, the area investigated at Pasadena includes Mono, Inyo, Tulare, Kings, and San Luis Obispo Counties and all California counties to the south.

dred kilometers distant are usually clearly recorded, and the seismograms have contributed materially to the study of travel times at greater distances, up to 1000 km. Most locations of epicenters in southeastern California discussed in this paper depend in considerable measure on the time of beginning at Tucson.

Table 1 lists the stations which have furnished seismograms used in this paper. Pasadena is equipped with short and long period Benioff instruments recording all three components, one long period and two short period Wood-Anderson torsion seismographs, and other instruments; at Palomar there is one short period vertical Benioff seismograph recording on cinematographic film, which is projected on a screen for measurements (on the projection 1 second corresponds to 2 mm); all other stations of the Pasadena group are equipped with one short period Benioff vertical seismograph and two horizontal standard Wood-Anderson torsion instruments ($V = 2{,}800$; $T_0 = 0.8$ sec; $h = 0.8\pm$). The maximum magnification of the short period Benioff verticals is of the order of magnitude of 100,000 at the inland stations, depending on the unrest of the ground, and much less (10,000±) at the two coastal stations. All short period instruments of this and most of the following groups have a drum speed of about 1 mm/sec. Radio time signals are recorded automatically 6 times in 24 hours at all stations of the Pasadena group, and once or twice at nearly all the other stations of Table 1.

TABLE 1.—*List of stations*

Station	Elevation meters	Latitude north	Longitude west
Pasadena	295	34° 08.9′	118° 10.3′
Mount Wilson	1742±	34 13.5	118 03.4
Riverside	250±	33 59.6	117 22.5
Santa Barbara	100±	34 26.5	119 42.9
La Jolla	8	32 51.8	117 15.2
Haiwee	1100±	36 08.2	117 57.9
Tinemaha	1180±	37 05.7	118 15.5
Palomar	1700±	33 21.0	116 51.5
Berkeley	81	37 52.3	122 15.6
Stanford (Palo Alto)	83	37 25.1	122 10.8
San Francisco	100	37 46.4	122 27.2
Lick (Mt. Hamilton)	1282	37 20.4	121 38.6
Fresno	88	36 46.1	119 47.8
Boulder City	776	35 58.9	114 50.1
Boulder Dam	237	36 00.9	114 44.2
Overton	395	36 31.9	114 26.5
Pierce Ferry	417	36 07.3	114 00.3
Tucson	770	32 14.8	110 50.1

At Berkeley, there are two standard torsion seismographs, one short period vertical Benioff, three components of long period Galitzin instruments, and three old instruments with smaller magnification; at Fresno, there is one standard torsion seismograph, and at the other stations of the Berkeley group there are two components of this type. The Boulder City station is equipped with one standard torsion seismograph, the other stations of this group with various short period instruments having magnifications between 1000 and 8000. Finally, at Tucson there are two torsion instruments ($V = 430$, $T_0 = 10$ sec) besides the two Benioff vertical instruments mentioned.

All stations of Table 1 are marked by triangles on Plate 1 [2] except Tucson which is about 3° east of the lower right corner of the map.

PROCEDURE IN LOCATING EPICENTERS

To locate the epicenters of the shocks involved, as a first step the method used in an earlier paper (Gutenberg, 1932a) was employed. Pairs of stations were selected which had either approximately the same arrival times for $\bar{P}$, the wave having its deepest point in the layer underlying the sediments ("granitic layer"), or for Pn, the wave traveling below the "Mohorovičić discontinuity" (in the "ultrabasic material"; *see* Gutenberg, 1943). The first type of wave has an apparent velocity of about 5.58 km/sec, the second type of about 8.06 km/sec. If the $\bar{P}$-wave arrives at one station x seconds earlier, or the Pn-wave y seconds earlier, than at another the first station will be approximately $5.58x$ km or $8.06y$ km closer to the epicenter than the second. On a map the locations of the stations are marked as accurately as possible, the approximate location of the epicenter (found in preliminary study) is plotted, and also pairs of points, having the same distance from this approximate epicenter (one of these points is the nearer station, just mentioned, the other a point near the farther station but closer to the assumed epicenters by $5.58x$ or $8.06y$). The epicenter which fits the data must be on lines having the same distance from each such pair of points. By plotting all these lines, a second approximation to the epicenter can be found; if this differs too much from the first, the method is repeated.

To obtain higher accuracy, a method was used which had been employed by Gutenberg and Richter (1937) in the determination of the epicenters of teleseisms: the residuals were calculated, assuming the second approximation of the epicenter and the given velocities, and plotted against the azimuth of the stations at the epicenter. These residuals must lie approximately on a sinusoidal curve. Usually, the

[2] All illustrations were drafted by Mr. John M. Nordquist.

results for $\overline{P}$- and *Pn*-waves were plotted on the same figure with different symbols (Fig 1a) since the velocities of the two waves are different, and, therefore, a given residual in time corresponds to an error in distance which is different for $\overline{P}$ and *Pn*. There is an advantage in plotting the time residuals and not the residuals in distance, since it is then easier to adjust the origin time if the sinusoidal curve is asymmetric to the zero line. Table 2 illustrates the method.

TABLE 2.—*Earthquake No. 22 (Table 3)*

Second approximation of epicenter: 33°48′ N., 118°14′ W.

Station	Distance km Δ	Azimuth degr.	Time $\overline{P}$ sec.	Δ/5.58	Difference	Residual sec.	Time *Pn* sec.	Δ/8.06	Difference	Residual sec.
Riverside	82.2	75	52.0	14.7	37.3	+0.9	.....	.....	.....	.....
Palomar	137.1	110	1.2	24.6	36.6	+0.2	0.0	17.0	43.0	0.0
La Jolla	138.2	140	1.0	24.8	36.2	−0.2	59.9	17.2	42.7	−0.3
Santa Barbara	154.1	300	3.5	27.5	35.9	−0.5	2.1	.....	43.0	0.0
Haiwee	260.4	5	23.4	46.7	36.7	+0.3	.....	.....	.....	.....
Fresno	358.7	340	40.6	64.3	36.3	−0.1	.....	.....	.....	.....
Tinemaha	365.6	0	41.3	65.5	35.8	−0.6	.....	.....	.....	.....
Boulder City	393.8	135	.....	.....	.....	.....	32	48.9	43.1±	+0.1±
Tucson	712.5	100	.....	.....	.....	.....	11¼	28.4	43¼	+¼
Average	.....	.....	.....	.....	36.4	.....	.....	.....	43.0	.....

All calculations of distances were made by using Richter's (1942a) method and tables. This saved many days of calculation.

The azimuths used in the calculation are taken from the map on which the second approximation for the epicenter was determined. In general, it is sufficient to measure them to the nearest 10°. As in Table 2, readings of $\overline{P}$ at stations closer than 60 km from the epicenter were omitted to avoid larger errors introduced by uncertainty in the depth of focus which will be discussed later. On the other hand, *Pn* was not used where a wave path under the Sierra Nevada might have introduced a delay (in Table 2, at stations Haiwee and Tinemaha). The residuals for $\overline{P}$ and *Pn* in Table 2 are plotted in Figure 1a as a function of the azimuth. The resulting sinusoidal curve is practically symmetrical to the zero line, indicating that the averages used (last line of Table 2) are a good approximation. If most of the residuals are either on the part of the sinusoidal curve below or above the zero line, the average does not correspond to the true zero; the latter can then be found from the plot. For stations at relatively short distances, the travel times for $\overline{P}$ must be corrected for the effect of focal depth to find the origin time of the shock; otherwise the average corresponding to the final zero line

for the $\bar{P}$-curve gives the origin time of the shock directly, while the average of $Pn - \Delta/8.06$ gives the intercept time of Pn. Other phases may be treated similarly.

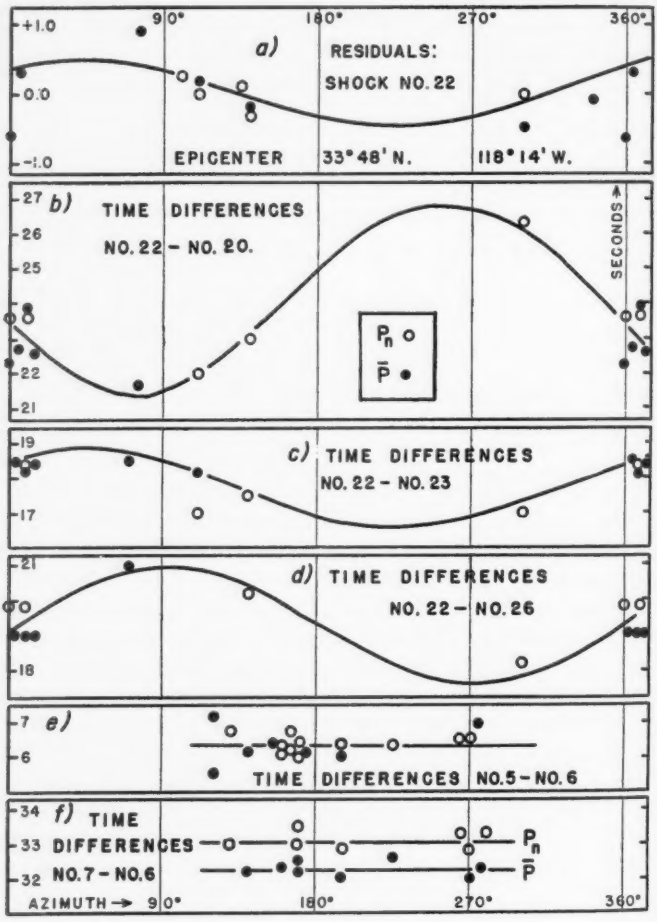

FIGURE 1.—*Diagrams used in location of epicenters*
For details, *see* text.

The shift of the epicenter necessary to obtain better agreement with the data can be found directly from the sinusoidal curve by considering the deviations of the sinusoidal curve from the zero line at the azimuths

90° and 180°. In Figure 1a, these deviations are about +0.4 and −0.3 second, respectively. This means that the correct epicenter is farther west and south than that assumed. The amount of the shift is given by the time difference multiplied by the velocity. Assuming an average velocity of 6½ km/sec for $\overline{P}$ and *Pn* we find that in our example a shift of about 3 km west and about 2 km south is required. This corresponds to a shift of about 2′ (1′ = 1.5 km) west and about 1′ (= 1.8 km) south. Thus, the corrected co-ordinates of the shock are 33°47′ N., 118°16′ W. (In case of larger corrections, $\overline{P}$ and *Pn* must be treated separately with their proper velocities.) The origin time is approximately $0^h41^m36.6^s$. The intercept time of the *Pn*-waves is 43.0 − 36.6 = 6.4 seconds. All origin times were revised later after the velocities of the waves had been redetermined, and the effect of the depth of focus was considered for each station. The final corrections did not exceed 0.3 second for 42 out of the 50 shocks.

If data are available for shocks with epicenters not far apart, the accuracy in the determination of the co-ordinates can be slightly increased by plotting time differences between the shocks. In Figure 1b-f examples are given. We find from Figure 1b that shock no. 20 originated about 10′ west and 3′ south of no. 22; from Figure 1c that the epicenter of no. 23 was about 2′ north and 3′ east of no. 22; and from Figure 1d that no. 26 occurred about 5′ due east of no. 23. This affords not only a good picture of the relative location of the several epicenters, but also a final solution for all epicenters involved in one group can be obtained. This must give the best solution not only for the relative location of the epicenters involved, found from comparisons, but also for each shock treated individually. Usually, the final adjustments are small and can be found directly by writing down the corrections required for each shock from all the results involved. In this way, the final co-ordinates of shock no. 22 were found by combining the relative and individual locations of all shock nos. 20-27 with the results obtained above for this shock, as follows: 30°47′ N., 118°15′ W.

Figure 1e illustrates two shocks originating practically at the same focus. Both occurred in northern Owens Valley (Long Valley) on September 14, 1941, about 18 minutes apart. Figure 1f shows that the shock on December 30, 1941, originated within the same area; the fact that the arrival of the *Pn*-waves was delayed as compared with that of $\overline{P}$-waves, but that each set of differences is approximately on a straight line, indicates that the epicenters of the two shocks were nearly the same but that the shock in December originated at a shallower depth. If t is the difference in time of the two straight lines in Figure 1f, a and b

the velocity of *Pn* and $\overline{P}$ respectively, the difference in depth, z, is given by

$$z = \frac{t}{\sqrt{\frac{1}{b^2} - \frac{1}{a^2}}} \qquad (1)$$

If $a = 8.06$, $b = 5.58$ km/sec, then $z = 7.7t$; a delay of *Pn* by 0.1 second relative to $\overline{P}$ can be produced by a decrease in focal depth by about ¾ km. Taking in our example $t = 0.7$ second, we find $z = 5$ km. The phases through the intermediate layers, too, show a delay in the second shock. If the velocities of such waves are 6 km/sec (*Py*) or 7 km/sec (*Pm*), the resulting values of z are $14t$ and $9.1t$, respectively. *Py* arrived about 0.3 second later (relative to $\overline{P}$) in the December shock; this gives $z = 4$ km, but the accuracy in using *Py* is noticeably less than in using *Pn*. It follows that the December shock originated at a depth of about 5 km shallower than those in September. The example shows that a delay of about ½ second is the least which can be ascertained beyond doubt in well-recorded shocks and that the method here used permits the finding of differences in focal depth of shocks from the same epicenter when the difference is at least 4 km. For epicenters separated by more than a few kilometers, the accuracy in finding the difference in focal depth is considerably less.

RESULTS FOR EARTHQUAKES 1933–1941

The methods described in the preceding section were employed with records of 50 earthquakes, including most of the larger shocks in southern California (1933-1941). Fewer are used for the years 1933-1937 than for 1938-1941, due mainly to the fact that in the later years some of the improvements mentioned in the Introduction made it possible to use smaller shocks in outlying regions. Even now, shocks in southern Imperial Valley and most of eastern California must have a magnitude of at least 5½ to yield reliable results. For example, a shock on March 2, 1942, near 115.8° W., 34.0° N., Magnitude 5.1, with *Pn* recorded at 12 stations with epicentral distances between 114 and 498 km, could not be located with the same accuracy as some of the shocks with Magnitude 4 closer to the Pasadena-Mt. Wilson-Riverside region because most of the stations had only small differences in azimuth; the exact determination of the origin time offered difficulties, as too few $\overline{P}$-phases could be ascertained. For shocks in the Sierra Nevada, the data available need to be more comprehensive than in other districts as some can be used only for the determination of the effect of the root of the Sierra.

Except for shock no. 26, no aftershocks of the Long Beach earthquake (no. 27) were investigated, as it is expected that these shocks will be dealt with by H. O. Wood and C. F. Richter. On the other hand, some small aftershocks of other earthquakes were included for comparison with the corresponding main shock.

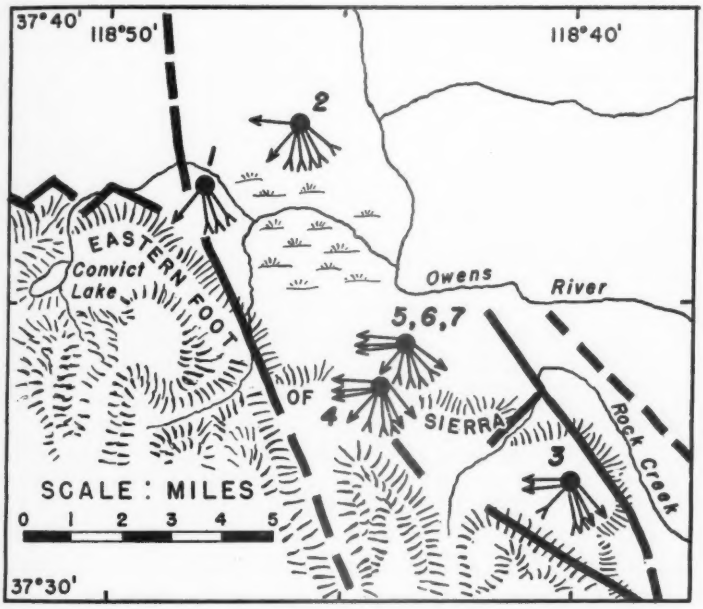

FIGURE 2.—*Map of epicenters in Northern Owens Valley (Long Valley)*
Fault traces after Mayo (1941). Epicenter 4 same as 5-7; it was plotted separately to show slightly different distribution of compressions and dilatations.

The principal data for the shocks are tabulated in Table 3 in which the column "number of stations" includes only those at which body waves were recorded. The numbers were assigned roughly from north to south; the epicenters are plotted in Plate 1. The "magnitude" of the shocks is discussed in the next section. Origin times given in Table 3 are based on the revised value for the velocity of $\overline{P}$ (5.577 km/sec) to be discussed later.

Plate 1 confirms previous findings that most of the larger shocks occur in a few small areas near major faults. This contrasts with the results for smaller shocks. "Most of the many small shocks are located close to one or another of the numerous minor faults which are common

throughout the region. Only the larger shocks show definite association with the larger fractures" (Gutenberg and Richter, 1941, p. 99).

Figures 2 to 4 show in greater detail the location of the epicenters in three areas. As already pointed out, the relative location of the epicenters is more accurate than the absolute. In all figures the direction

TABLE 3.—*Data for shocks*

No.	Date	Origin time hr. min. sec.	Longitude deg. min.	Latitude deg. min.	Magnitude	No. of stations	Range of distances km
1	1940, July 8	02 57 36.5	118 48	37 37	4.8	13	76–941
2	1940, July 22	15 00 32.9	118 46	37 38	4.6	12	75–940
3	1938, Dec. 3	09 42 50.2	118 40	37 32	5.5	14	61–926
4	1941, Sept. 14	08 43 31.8	118 44	37 34	5.8	16	67–932
5	1941, Sept. 14	10 21 18.7	118 44	37 34	5.5	16	67–932
6	1941, Sept. 14	10 39 11.9	118 44	37 34	6.0	16	67–932
7	1941, Dec. 30	22 48 44.0	118 44	37 34	5.4	16	67–932
8	1939, Jan. 7	12 21 49¼	117 45	36 00	5.0	13	25–762
9	1938, Sept. 17	06 23 04.7	117 41	35 39	4.8	14	59–738
10	1939, May 4	12 44 46¾	114 49	35 58	5.0	9	2–552
11	1938, Aug. 17	23 39 44¼	116 13	34 52	4.5	9	143–576
12	1939, March 22	11 16 36½	116 08	34 42	3.8	9	139–570
13	1939, Feb. 23	00 45 50.6	119 01	34 52	4.6	13	80–814
14	1939, Feb. 23	01 18 45.7	119 00	34 54	4.8	13	83–813
15	1939, May 7	18 48 03.7	119 03	34 54	4.4	12	79–818
16	1941, Sept. 21	11 53 07.2	118 56	34 52	5.2	16	86–806
17	1941, June 30	23 50 54.8	119 35	34 22	5.9	18	14–849
18	1935, July 13	02 54 16.5	117 54	34 12	4.7	11	15–693
19	1941, Jan. 21	17 34 46.9	118 03	33 58	4.1	7	23–348
20	1940, Oct. 10	21 57 12.3	118 27	33 46	4.7	11	49–731
21	1938, Aug. 30	19 18 12.0	118 16	33 44	4.4	9	47–714
22	1941, Nov. 14	00 41 36.3	118 15	33 47	5.4	13	41–714
23	1941, Oct. 21	22 57 18.5	118 13	33 49	4.9	11	37–711
24	1941, Oct. 22	02 32 21¾	118 13	33 52	3.8	8	31–370
25	1939, Dec. 27	11 28 49.0	118 12	33 47	4.7	10	41–709
26	1933, Oct. 2	01 10 17.6	118 08	33 47	5.4	12	41–696
27	1933, Mar. 10	17 54 07.8	117 58	33 37	6.3	12	62–684
28	1940, Apr. 18	10 43 43.9	117 21	34 02	4.4	11	5–640
29	1939, Nov. 7	10 52 08.4	117 17	34 00	4.7	8	9–633
30	1940, Feb. 19	04 06 55.7	117 03	34 01	4.6	11	30–612
31	1935, Oct. 24	06 48 07.6	116 48	34 06	5.1	7	54–358
32	1938, May 31	00 33 54.0	117 32	33 40	5.3	12	39–646
33	1938, July 5	10 06 54.6	117 35	33 40	4.4	10	40–650
34	1938, Aug. 6	14 00 55.0	117 34	33 42	3.8	9	36–650
35	1940, Feb. 25	13 23 47.1	117 32	33 41	3.4	5	37– 94

TABLE 3.—*Data for shocks—Continued*

No.	Date	Origin time hr. min. sec	Longitude deg. min.	Latitude deg. min.	Magnitude	No. of stations	Range of distances km
36	1940, May 17	21 03 58.5	116 18	34 05	5.4	14	96–682
37	1940, May 17	21 51 20¼	116 20	34 04	5.2	13	97–680
38	1940, May 17	22 04 30.6	116 19	34 04	4.6	13	98–682
39	1940, May 17	23 21 32.7	116 20	34 04	5.0	13	97–680
40	1940, May 31	21 27 01.2	116 20	34 05	4.7	10	95–552
41	1940, May 31	22 54 28.0	116 20	34 06	4.3	9	97–552
42	1940, June 1	22 13 10.2	116 20	34 05	4.5	10	95–551
43	1940, June 6	14 21 15.1	116 19	34 00	4.3	10	88–547
44	1940, June 6	15 56 37¼	116 22	34 01	4.4	10	87–552
45	1940, June 11	11 51 18.1	116 19	34 02	4.4	10	91–549
46	1937, Mar. 25	08 49 03.0	116 28	33 28	6.0	8	99–544
47	1939, May 12	11 25 02¼	116 26	33 28	4.5	9	101–541
48	1940, June 4	02 35 08.3	116 26	33 00	5.1	10	55–532
49	1940, July 21	00 36 03.0	115 59	33 10	4.4	10	83–493
50	1940, May 18	20 36 40.9	115 30	32 44	6.7	15	144–844

of the first motion at the stations has been indicated by arrows, either pointing toward the station, indicating a compression, or toward the epicenter, indicating that the first movement at a station in this direction was a dilatation. The findings in all 50 shocks confirm previous results of the author (Gutenberg, 1941) and in several instances remove discrepancies resulting from incorrect location of the epicenter. Conclusions from these arrows require familiarity with the theory and should not be drawn without consulting the original paper.

Figure 2 shows the location of the epicenters 1-7. Faults are drawn after Mayo (1941). The arrows indicate that in all shocks the faulting occurred in the way usual for California; the western block moved northward relative to the eastern block. The small differences between the shocks may be due either to slightly different directions of the faults or to relatively small vertical movements superimposed upon the horizontal shifts. From routine location work, shock nos. 1-3 had been attributed previously to the epicenters *X*23, *X*20, and *X*18 respectively (Gutenberg, 1941, Fig. 4; Table 2).

In Figure 3, the epicenters 13-16 are plotted on a larger scale. The results leave little doubt that they are on different branches of the fault systems in this area, though the absolute locations may be in error by as much as 3 miles (5 km). The compressions and dilations in these shocks show different patterns, something which is to be expected in this region with faults in various directions and faulting of all kinds. Additional and more accurate data are needed for trustworthy interpretation.

The compressions and dilatations in shock nos. 20-26 (Fig. 4) show a picture more similar to that in Figure 2. The conditions in shock no. 26 have been discussed in detail previously (Gutenberg, 1941, p. 292, epicenter *I*.23). The pattern for the shock nos. 22, 23, and 24 is similar.

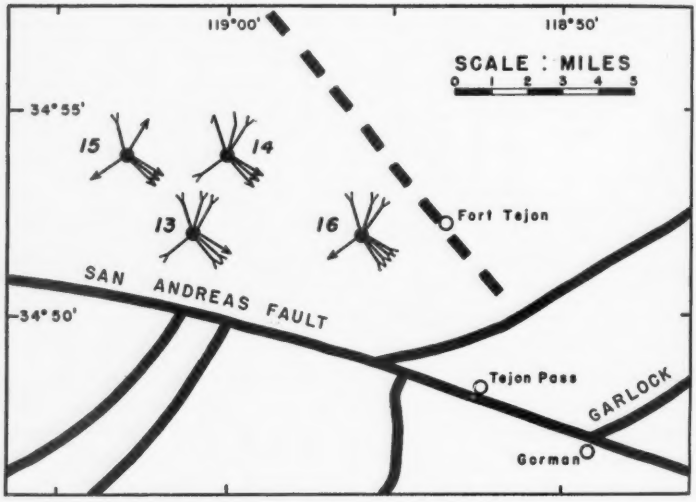

FIGURE 3.—*Map of epicenters near intersection of San Andreas and Garlock faults*

Fault traces after Jenkins (1938) and Willis (1923).

In all of them the compressions in the northeast sector extend somewhat to the north beyond the line perpendicular to the fault, contrary to the conditions in shock nos. 20, 21, and 25, which correspond to the usual pattern. A small vertical movement or an effect of the thick sedimentary layers to the north of the epicenters may account for the difference in the first four shocks. In general, shocks in the area of Figure 4 have indicated the usual movement.

Again Figure 4 indicates that most of these shocks shown have originated on different branches of the fault system. Figures 2-4 and the shocks in other groups, not plotted in detail, indicate that wherever data are available successive shocks are distributed over subparallel branches of the fault systems as well as along the larger fractures; reference should be made here also to shock no. 7, which, according to our findings in the preceding section, originated at about the same epicenter as shock nos. 4-6, but at a different depth.

Shock no. 19 has been investigated by C. F. Richter (1942).

Shock no. 28 had an epicenter only a few kilometers from a quarry near Colton. Records of blasts from this source have been recorded on several occasions over a term of years and have been studied by C. F. Richter (unpublished) who has found a velocity of about 6 km/sec

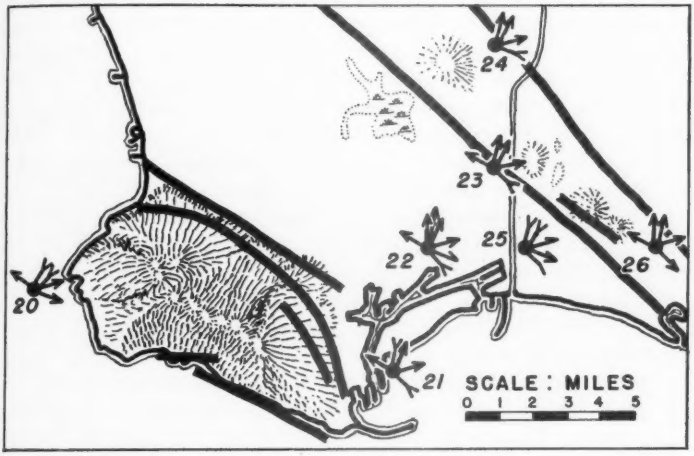

FIGURE 4.—*Map of epicenters northwest of Long Beach*
Fault traces after Willis (1923).

for the first waves. Notwithstanding, shock no. 28 cannot be located by assuming such a high velocity without obtaining residuals between the observed and calculated time which exceed the limits of error. A velocity of 5.59 km/sec for $\overline{P}$ gives the best agreement between observations and calculated values. A least square solution for shock nos. 28 to 31, taking the velocity of $\overline{P}$ and the origin times as unknown quantities, gives (5.58 $\pm$ 0.02) km/sec for the velocity of $\overline{P}$. All observations could be explained by assuming that the "granitic layer" with the usual velocity of 5.58 km/sec is covered by one in which the velocity is about 6.0 km/sec. There is no geological support for this suggestion; for details, *see* Gutenberg (1943).

Shock no. 30 had been located in the routine work about 0.5° farther southwest than the position adopted here (epicenter *H*.24, Gutenberg, 1941). In their revised new locations, the shocks of this group (28-31) show exactly the distribution of compressions and dilatations to be expected from the usual faulting in California, while with the original results on the epicenters some of the Riverside records showed a direction of the first impulse opposite to that expected at the assumed epicenter.

Shock nos. 32-35 (Santa Ana Mountains) have too early $\overline{S}$-phases. This, and a relatively large shaken area for number 32, led originally to a different epicenter and to the suspicion that these shocks originated at a depth of perhaps 30 km in an intermediate layer. However, the travel times could not be explained on this assumption, and the P-phases fit well with those for normal depth. The shocks occurred in a region with small population, and the maximum intensity in no. 32 may easily have been greater than the grade 6 which is indicated by the observations. However the granitic layer in this region possibly is slightly deeper than usual, say by 5 km. In the records at Pasadena and Mount Wilson, small deviations from normal are expected because the waves reaching these stations have to pass through rather deep sedimentary layers. Shocks from other areas with paths through these layers, for example from the Inglewood fault to Pasadena, Mount Wilson, and Riverside, or from Imperial Valley to the first two stations, or from the San Bernardino Mountains to Pasadena, develop unexpectedly small amplitudes in the first recorded wave, or complicated impulses, or arrive from a direction deviating by a noticeable amount from the direction towards the epicenter. The early arrival of $\overline{S}$ will be discussed in a later section.

Shock no. 46 is the "Terwilliger Valley" earthquake investigated by Wood (1937). The revised epicenter is 7′ (about 6 miles) farther east; the difference is mainly due to an earlier reading for the beginning at Tucson (08:50:17 instead of 08:50:21); this beginning (on a torsion seismograph with a free period of 10 seconds) is small but definite. For shock no. 47 from practically the same epicenter, the record of the short period Benioff vertical seismograph at Tucson leaves no doubt about the beginning.

Shock no. 50 is the "Imperial Valley earthquake". The results found previously for the distribution of compressions and dilatations (Gutenberg, 1941, p. 290, epicenter $B.10$) were confirmed by additional records.

DEPTH OF FOCUS

The most accurate determinations of focal depth are furnished by the times of arrival of the P- and S-phases at short epicentral distances. If Δ = epicentral distance, D = hypocentral distance, h = focal depth, O = origin time, V and v the velocities of $\overline{P}$ and $\overline{S}$ respectively, $T = P - O$ and $t = S - O$ their respective travel times, $a = t - T = \overline{S} - \overline{P}$, then we have

$$h^2 = D^2 - \Delta^2 \tag{2}$$

$$D = VT = vt \tag{3}$$

$$D = \frac{a\,v\,V}{(V - v)} \tag{4}$$

In California, $V = 5.58$ km/sec, $v = 3.24$ km/sec, $D = 7.7\ a$.

Usually, the travel times T of $\overline{P}$ are found much more accurately than the travel times t of $\overline{S}$; on the other hand, both depend on the adopted origin time, whereas their difference a does not. For this reason, equation (3) was applied only to T, and, in addition, equation (4) was used. Both furnish values for D, and then equation (2) gives h, supposing that Δ is known accurately. Errors in the determination of Δ, which depends on the adopted epicenter, affect least the results for very small epicentral distances. Consequently, most weight was given to results based on seismograms from distances less than 25 km. (Table 4.) The findings based on other seismograms confirm these figures, but the errors increase rather rapidly with distance.

TABLE 4.—*Calculated depths of focus*

Shocks numbered as in Table 3. For smybols, *see* text.

Shock No.	Station	Δ km	$T=P-O$ sec.	$a=S-P$ sec.	Calculated depth in km.		
					From T	From a	Average
10	Boulder City.....	2	1.8±	2±	10±	15±	13±
17	Santa Barbara.....	14	5.5	3.3	27	21	24
18	Mount Wilson.....	15	4.9	?	23	.....	23
19	Pasadena.....	23	4.9	?	15	.....	15
28	Riverside.....	5	3.5	2.2	19	16	17
29	Riverside.....	9	5.0	2.6	26	18	22

The data for shock no. 10 (near Boulder City, Nevada) are less certain than those for the others, since its origin time is less accurate, and a had to be taken from aftershocks. The average of all data is about 19 km. On the other hand, the average thickness d of the granitic layer in the region of the shocks is found (Gutenberg, 1943, equation 3) from

$$d = \frac{\tau}{2\sqrt{\left(\frac{1}{v_1}\right)^2 - \left(\frac{1}{v_2}\right)^2}} + \frac{h}{2} = 7.17\tau + \frac{1}{2}h$$

where τ is the intercept time of the wave through the second layer, about 1.29 seconds (*see* the section on "other P-phases" in this paper), $v_1 = 5.577$, the velocity in the granitic layer, and $v_2 = 6.053$, the velocity beneath this layer. This gives $d = 9.2 + 9.5 = 18.7$ km. Since the source of the shocks cannot be below the layer with a velocity of 5.577 km/sec, as otherwise $\overline{P}$ could not be recorded, it follows that h cannot exceed d; h and d are both about 18½ km. This agrees with the fact that the travel time curves of Py (to be discussed later) and $\overline{P}$ are practically tangent to each other. Foci of earthquakes have been found

frequently in the discontinuity between the granitic and the next deeper layer where the breaking strength is likely to be less than in the interior of each layer.

The ratio between the maximum intensity and the radius of perceptibility of the various shocks indicates approximately the same focal depth with few exceptions, these mainly from epicenters in sparsely settled areas (Gutenberg and Richter, 1942).

MAGNITUDE OF SHOCKS

In order to obtain a scale which indicates the relative energy of earthquakes and permits the determination of the scale value for a given shock from instrumental data, Richter (1935) has discussed what he termed "magnitude" of a shock. He defined the magnitude as the logarithm of the maximum trace amplitude, expressed in microns, with which the standard short-period torsion seismometer would register that shock at an epicentral distance of 100 kilometers. Tables (chiefly Richter, 1935, Table 1) and a nomogram (Gutenberg and Richter, 1942, Fig. 1) have been provided to aid in the calculation of the magnitude, M, for shocks in southern California originating at a depth of about 18 km; these can also be used for shocks occurring in other regions at such depths and with such structures as are not too different from those in southern California. In the present paper, the nomogram just mentioned was used. It was constructed by John M. Nordquist and is based chiefly on the table given by Richter (1935). The effect of a different focal depth is small, so long as the source of the shock is in the (granitic) layer below the sediments, and the stations are not too close to the source. For shocks at the usual depth of about 18 km in southern California, Gutenberg and Richter (1942) have found the following average relationship between the magnitude M and the maximum intensity I (Modified Mercalli intensity scale of 1931):

M	3½	4	4½	5	5½	6	6½	7
I	3.7	4.5	5.3	6.2	7.0	7.8	8.7	9.5

The value of M calculated from the records of short-period torsion instruments at a given station is influenced by the deviation of these instruments from the standard ($T_0 = 0.8$ sec, $V = 2800$, $h = 0.8$) and by the effects of the local structure on the arriving waves. (For a summary on "ground factors", *see*, for example, Gutenberg, 1932b, p. 193, 259.) Richter (1935, p. 12) has given mean corrections for the stations in southern California from records 1929-1931. These values were used in calculating the magnitudes (average of all stations) for the shocks in Table 3. The deviations from the average magnitudes furnished a

new mean correction for each station, now including those in northern California and Boulder City. With these revised corrections, all magnitudes were recalculated (Table 3). The latest magnitude corrections for the stations with standard torsion seismographs are given in Table 5. The findings of Richter (1935) for the earlier years do not differ by

TABLE 5.—*Corrections to be added to calculated magnitude at various stations*

Pasadena	+0.2	Mt. Wilson	0.0	Riverside	+0.2
Santa Barbara	−0.1	La Jolla	0.0	Haiwee	0.0
Tinemaha	−0.2	Berkeley	+0.2	San Francisco	+0.1
Lick (Mt. Hamilton)	+0.1	Stanford	−0.1	Fresno	−0.2
Boulder City	0.0				

more than 0.1 at any station. Even these small differences may depend partly on the prevailing azimuths of the shocks from the station, since mountain ranges or deep fault zones between the station and the source sometimes reduce the amplitudes below average. (*See* Gutenberg, 1941, p. 291, 295.)

If the mean corrections given in Table 4 are applied in each case, the remaining residuals give the effect of the wave paths just mentioned

TABLE 6.—*Average residuals in magnitude at different stations for groups of shocks (numbers as in Table 3) in various parts of southern California*

Shock No.	1–7	13–16	20–27	28–31	32–35	36–45	48–50
Pasadena	+0.3	0.0	0.0	0.0	−0.2	0.0	−0.1
Mt. Wilson	0.0	0.0	−0.1	0.0	−0.1	0.0	0.0
Riverside	+0.1	0.0	+0.1	.....	.....	−0.2	+0.1
La Jolla	+0.1	−0.1	−0.1	+0.2	0.0	0.0	.....
Santa Barbara	+0.2	−0.1	0.0	−0.2	−0.1	0.0	−0.1
Haiwee	+0.1	0.0	−0.1	0.0	+0.1	+0.1	0.0
Tinemaha	.....	+0.1	0.0	0.0	−0.1	+0.2	−0.1
Fresno	.....	0.0	.....	0.0	−0.1	+0.2	−0.1
Lick	−0.2	+0.1	+0.1	.....	0.0	+0.2	+0.1
Stanford	−0.2	0.0	+0.2	.....	.....	0.0	+0.2
Berkeley	−0.1	+0.2	+0.2	.....	0.0	0.0	+0.2
Boulder City	−0.1	+0.1	+0.1	+0.1	0.0	−0.2	0.0

and errors in the tables or graphs used in the calculation of M for the distance in each specific case. In a very few cases, such final residuals amounting to ± 0.4 were found. The great majority did not exceed ± 0.2. Average residuals for shocks in various parts of southern California are given in Table 6.

Each figure of Table 6 corresponds to a certain small range of distances. In plotting these figures against distance, one finds the correc-

tions to be applied to Richter's (1935) original Table 1 or the nomogram used in this paper. These residuals are surprisingly small and indicate that Richter's table was accurate within 0.1 unit of the magnitude scale. The maximum residual in Table 6 is the first, indicating that the largest waves of the seven shocks in northern Owens Valley (Long Valley) gave a magnitude 0.3 unit too large from the Pasadena records after applying the general correction of +0.2 at Pasadena. Since this corresponds to the logarithm of the amplitudes, it follows that these shocks showed at Pasadena average maximum amplitudes about twice as great as expected from the average at all stations. However, the table shows that the residuals in the first column, though small, are systematically positive at the stations south of the epicenter, while negative at the Berkeley group. This could be explained by assuming that a fraction of the energy in the maxima (probably *S*-waves) was lost in passing through or under the Sierra Nevada. The true magnitude of these shocks probably corresponds more closely to the average of the southern stations and consequently is more likely to be 0.1 or 0.2 unit higher. The residuals at the Berkeley group would then be about —0.3. In addition to the usual decrease of the amplitudes with increasing distance, about half the amplitudes had been diverted into other directions or absorbed along the paths to the Berkeley group. More data are required for the investigation of this problem and the question as to how far there are other systematic residuals.

ACCURACY IN TIME DETERMINATIONS AND LOCAL EFFECTS ON ARRIVAL TIMES OF WAVES

The observed arrival times of waves are influenced by the effect of the local conditions under the stations as well as by errors, such as inaccurate correction of the clock making the time marks on the record, uneven speed of the recording drum, and errors in reading the time of a phase on the seismogram. Estimates on their combined effects have been made since the early days of recording earthquakes; in most instances the probable error of the results has been underestimated. To obtain more reliable data, the following procedure was followed: For all shocks of magnitude 4 or more, the observed time differences of *Pn* between Pasadena and Mount Wilson and between Pasadena and Riverside were tabulated, as well as the time differences of $\overline{P}$ at Pasadena and Mount Wilson from distances beyond 100 km. It was assumed that the apparent velocity of *Pn* is 8.06 km/sec, that for $\overline{P}$, 5.58 km/sec, and the time intervals were calculated. An error in the assumed velocities affects the calculated time differences between Pasadena and Mount Wilson by less than 0.1 second for a certainty since the difference in epicentral

distance between Pasadena and Riverside was less than 76 km, whence the error in the calculated time difference probably remained below 0.1 second, for the velocity of *Pn* in the region is now known probably within 0.05 km/sec.

TABLE 7.—*Calculated residuals in time differences (seconds)*

Minus sign indicates that time at Pasadena is too early relative to other station

Shock No.	Pasadena-Mt. Wilson		Pasadena-Riverside	Shock No.	Pasadena-Mt. Wilson		Pasadena-Riverside
	Pn	$\overline{P}$	*Pn*		*Pn*	$\overline{P}$	*Pn*
1	−0.3	0.0	+0.4	36	−0.2	+0.7	.....
2	−0.4	.....	+0.2	37	−0.3	−0.3	.....
3	0.0	(+0.4)	−0.1	38	−0.2	.....	.....
4	−0.4	0.0	+0.1	39	−0.2	+0.1	.....
5	−0.5	+0.3	+0.1	40	−0.1	−0.4	.....
6	−0.7	0.0	−0.3	41	−0.2	.....	.....
7	−0.2	+0.3	.....	42	0.0	−0.1	.....
8	−0.4	.....	+0.1	43	+0.3	.....	.....
9	0.0	+0.8	+0.1	44	+0.3	.....	.....
10	−0.2	.....	+0.4	45	+0.1	.....	.....
11	−0.4	.....	+0.2	46	+0.1	+0.2	.....
17	−0.4	−0.1	+0.2	48	.....	0.0	0.0
31	−0.5	−0.5	.....	50	+0.1	+0.6	0.0

The residuals found in this way are given in Table 7. Shock nos. 1-17 are in general to the northward from the three stations, numbers 36-50 to the southward; number 31 is to the east of Pasadena but too close to Riverside to be used. Most of the epicenters 18-35 are less than 120 km from at least two of the three stations, and *Pn* does not exist or arrives later than $\overline{P}$ or P^*, while at these distances the effect of the focal depth influences the apparent velocity of $\overline{P}$ too much to permit guarantee of the accuracy required for our purpose.

The data of Table 7 indicate that the residuals for the differences of arrival of *Pn* at Pasadena and Mount Wilson depend on the azimuth of the epicenter. In shock nos. 1-17, all *Pn*-phases arrived too late at Mount Wilson, relative to Pasadena, whereas the residuals for numbers 36-50 are closer to zero and partly positive. That the last five shocks show positive residuals is at least partly a coincidence, since the epicenters of numbers 43-45 are practically the same as those of 36-42. The differences for $\overline{P}$ at Pasadena and Mount Wilson and the differences for *Pn* at Pasadena and Riverside do not show any definite influence of the location of the epicenter. Consequently, they were treated together, whereas the data for Mount Wilson-Pasadena were divided into

two groups; shock no. 31 was used only in the average. The result is shown in Table 8. (For data available, *see* Table 7.)

The results show that *Pn* arrives slightly late at Mount Wilson relative to Pasadena. This is to be expected because of the difference in elevation (1447 meters). For *Pn*, the effect of this elevation is given approximately by equation (1): $t = z/7.7 = 0.19$ sec; for $\overline{P}$, there

TABLE 8.—*Mean time differences between pairs of stations*

Stations	Phase	Shock No.	Time difference Seconds
Pasadena-Mt. Wilson	*Pn*	1–17	-0.33 ± 0.04
Pasadena-Mt. Wilson	*Pn*	36–50	-0.03 ± 0.04
Pasadena-Mt. Wilson	*Pn*	1–50	-0.19 ± 0.03
Pasadena-Riverside	*Pn*	1–50	$+0.11 \pm 0.04$
Pasadena-Mt. Wilson	$\overline{P}$	1–50	$+0.09 \pm 0.06$

Minus sign means that Pasadena was early relative to the other station; figure following sign "±" gives probable error; for a large body of data error should exceed this figure as often as fall below it.

should be no noticeable delay at long distances. In both waves, an additional delay will occur if the mountain slope is so steep as to interfere with the wave path. On the other hand, in the area south of Pasadena there are thick sedimentary layers, while to the north sediments are practically lacking near the station. A fault separating the granite and the sediments is about ¼ kilometers south of the station. The combination of all these elements explains the time differences of *Pn* in Table 8 within the limits to be expected. The small delay of $\overline{P}$ at Pasadena may be due to the fact that these waves arrive at a rather large angle of incidence and therefore are more affected by the sediments; this probably affects the sharpness of the beginnings, too, and may produce a less clear beginning of $\overline{P}$ at Pasadena.

The figures show that the accuracy of timing is high enough to bring out local effects. The probable errors permit a more definite answer to the question as to how accurately the times are being determined. The two equations used for the calculation of the probable error F of the result (n observations) and f of one observation, are

$$F = \frac{0.6745\, e}{\sqrt{n(n-1)}} \tag{5}$$

$$f = \frac{0.6745\, e}{\sqrt{n-1}} \tag{6}$$

e is the square root of the sum of the squares of the errors. Equation (6) gives $f = 0.13$ to 0.14 for all series concerning *Pn*, and 0.24 for $\overline{P}$. As

these are differences of two readings with approximately the same accuracy, we have to divide the results by $\sqrt{2}$ to find the probable error of a reading at one station. This gives a probable error of 0.1 second for *Pn*, and slightly less than 0.2 second for $\overline{P}$. Theoretically, in a large body of data for *Pn* a given range of errors should be expected with the following probabilities:

error	$\geqq 0.1$	$\geqq 0.2$	$\geqq 0.3$	$\geqq 0.4$	$\geqq 0.5$	seconds
once in	2.0	5.6	23	143	1342	measurements

For $\overline{P}$ the values of the first line should be approximately doubled. In considering these figures, we should remember that they refer to conditions similar to those prevailing in the instances used in Table 7. These "errors" contain effects connected with the local geology, mentioned above, which are a function of the azimuth and the angle of incidence of the waves. The true errors consisting of the effect of the time correction of the clock, the errors in reading the seismograms, and the effect of irregularities in the drum speed combined, exceed 0.2 second in only a very small percentage of the cases in which the arrival time of *Pn* at distances of about 100-500 km has been determined from records of the three stations used (and probably similarly at all stations of the group in southern California). As times on the record and time corrections of the clocks could be estimated only to the nearest tenth of a second, the accuracy can be increased by an increase in drum speed.

PRELIMINARY DATA ON TRAVEL TIMES AND VELOCITIES

GENERAL STATEMENT

On all records available for the 50 shocks of Table 3 outstanding phases were measured, and their times were plotted against distance for most shocks. In addition the data for each group separated in Table 3 by horizontal lines were plotted together and, finally, two plots were made on a larger scale, one for shock nos. 1-7 (epicenter in northern Owens Valley) and another for numbers 8-50, the latter with a scale for the distances of ½ cm for 1 km, and for the times of 1 cm for 1 second. The study of these travel time plots left no doubt that their detailed investigation would require a long time, and it was decided to leave this for a later paper. As far as findings from this study are available they are given in the following discussion. Otherwise, the following results are only preliminary. For a discussion of the wave paths, *see* Gutenberg (1943).

Pn

Records from shocks in California at distances beyond about 120 km (depending on local conditions and depth of focus) begin with *Pn*, the

wave having its deepest point below the Mohorovičić discontinuity. The apparent velocity of this wave in all shocks with sufficient data was calculated using the method of least squares and was found to be at least 8.03 and not more than 8.10 km/sec, with an average of 8.06 ± 0.01 km/sec (236 independent instances). Supposing the depth of the Mohorovičić discontinuity to be about 40 km, the true velocity of this wave

TABLE 9.—*Intercept time τ for Pn in seconds*

Numbers of shocks as in Table 3.

No.	1	2	3	4	5	6	7	8	9	10	11	12	13
τ	9.0	8.9	7.7	9.0	8.5	8.8	9.7	6.4	6.1	7.3	7.3	5.8	6.5
No.	14	15	16	17	18	19	20	22	27	28	29	30	31
τ	6.1	6.2	6.4	6.7	6.0	5.9	6.4	6.5	6.8	6.5	6.0	5.9	5.9
No.	32	36	37	38	39	40	41	42	44	45	46	48	50
τ	7.1	7.0	7.4	7.3	7.2	7.0	7.3	6.8	7.2	7.1	6.6	5.9	6.0

is 8.00 km/sec, probably within 0.05 km/sec. Previously the author (Gutenberg, 1932a) had found a value of 7.94 for the *apparent* velocity. The difference of 0.12 km/sec is mainly due to the fact that in the earlier study the most distant station was frequently Tinemaha. As the waves to this station arrive with a delay from many epicenters, as we shall see, too small a velocity was found in such instances. Moreover, the use of short period Benioff Vertical instruments at several stations where previously only torsion instruments were available reduced the number of cases in which the time adopted for *Pn* was slightly too late because a small beginning of *Pn* was not recognized on the record.

Besides the apparent velocity, it is important to know the time interval, measured along the axis of time, between the origin time of the shock and the point at which the straight travel time curve for *Pn* intersects the time axis. This time interval has been called the "intercept time" by Byerly. In the previous investigation (Gutenberg, 1932a) it was found to be 5.8 seconds (average). Table 9 leaves no doubt that its value shows differences depending upon the region. It is noticeably largest in shock nos. 1-7, originating in northern Owens Valley (Long Valley); next in order come nos. 36-45 (Little San Bernardino Mountains), no. 10 (near Boulder City), no. 11 (eastern Mojave Desert) and no. 32 (Santa Ana Mountains). A great value for τ indicates either that the shock was shallower than usual or that the boundary of the layer in which the phase has its deepest point is at a greater depth than usual. (For conclusions based on these data, *see* Gutenberg (1943).)

To find the residuals for *Pn* at the various stations, for each shock its proper time of intercept must be used. In Table 10 such residuals are given for selected shocks, excluding the stations Haiwee and Tinemaha, for which all residuals are given in Table 11, and some stations for which only a limited amount of data was available.

TABLE 10.—*Residuals of Pn in selected shocks (seconds)*

	Shock number (Table 3)										
	3	6	16	17	22	27	28	32	36	46	50
Pasadena	−0.1	−0.6	.....	−0.2	.....	.....	.....	.....	−0.1	−0.2	−0.3
Mt. Wilson	−0.3	−0.1	.....	+0.1	.....	.....	.....	.....	0.0	−0.6	−0.6
Riverside	−0.0	−0.3	+0.3	−0.4	.....	.....	.....	.....	.....	.....	−0.3
Palomar	.....	+0.1	0.0	.....	−0.1	.....	.....	.....	.....	.....	−0.1
La Jolla	.....	0.0	−0.5	0.0	0.0	.....	+0.2	.....	0.0	.....	+0.7
Santa Barbara	.....	+0.6	.....	.....	+0.2	−0.1	0.0	−0.6	+0.3	+0.4	+0.3
Berkeley	+0.4	−0.4	−0.5	−0.2	.....	−0.3	.....	+0.3	−0.2	.....	+0.1
Boulder City	.....	−1.0±	+0.2	+0.2	+0.3±	.....	−0.5	.....	.....	.....	+0.3
Tucson	+0.5	+0.6	+0.4	+0.3	−0.3	+0.8±	+0.7	−0.4	−0.2	−0.3	−0.2

The data of Table 11 leave no doubt that there is a delay in the arrival of *Pn* at the Owens Valley stations in many instances, especially at Tinemaha. Similar delays had been found previously by Byerly and Wilson (1935) which Byerly (1938) considered an effect of the root of the Sierra Nevada. There are instances where a similar delay has been found from records at Fresno or the Berkeley group of stations in one or another of the 50 shocks under discussion in this paper. All

TABLE 11.—*Residuals of Pn at Haiwee (H) and Tinemaha (T) in seconds*

Shock numbers as in Table 3. Effect of elevation (0.2 second) is considered.

No.	*H*	*T*	No.	*H*	*T*	No.	*H*	*T*
8	.....	−1.1	22	+1.1	+1.0	38	+0.2	+1.3
9	.....	−0.2	23	+1.0	+1.8	39	+0.5	+1.1
10	−0.4	.....	25	+1.1	.....	40	0.0	−0.4
11	0.0	+0.4	26	+0.5	+0.3	41	+0.3	−0.1±
12	+0.4	+0.6	27	+0.8	+1.0	42	.....	+0.5
13	.....	(+2.2)	28	−0.3	−0.1	43	+0.6	+0.9±
14	.....	+1.6	29	−0.2	.....	44	+0.6	+0.4
15	0.0	+1.4	30	0.0	+0.4	45	+0.8±	0.0±
16	+1.7	+1.6	31	−0.2	0.0	46	+0.7±	+0.4±
17	+2.2	+1.7	32	+0.4	+0.3	47	+0.2	+0.5±
18	−0.2	+1.0	33	+0.6±	.....	48	−0.1	−0.2
19	−0.1	+0.3	36	−0.1	+0.5	50	−0.2	+0.2
20	+1.1	+1.2	37	+0.7	+1.1			

these data have been used by the author (Gutenberg, 1943) in a study on mountain roots.

$\overline{P}$

At epicentral distances less than about 90 km the seismograms in California begin with $\overline{P}$, the wave which has traveled solely in the "granitic" and sedimentary layers. At somewhat greater epicentral distances this phase is usually outstanding because of its rather large amplitude (*see*, for example, Gutenberg, 1943, Pls. 1, 2); at distances of a few hundred kilometers, its amplitude decreases, and beyond 600 km it is only rarely identified beyond doubt. The results based on the registration of this wave are slightly less certain than those for *Pn* because, unlike *Pn*, the travel times are influenced noticeably by the effect of the depth of focus by an amount depending on the hypocentral distance. For a shock at a focal depth of 18 km the delay in the arrival of $\overline{P}$ compared with that for a source at the surface is given approximately by the following values:

Distance	0	10	20	30	40	60	80	100	150	200 km
Delay	3.2	1.9	1.2	0.9	0.7	0.5	0.4	0.3	0.2	0.1 sec

In the determination of the velocities of the $\overline{P}$-waves and the residuals at the stations, distances less than 60 km were not used; for the larger distances the corrections given above were applied. First, data from regional groups similar to those indicated in Table 3 were combined and the method of least squares was applied to calculate corrections to the origin time and to the velocity of $\overline{P}$. In this way, velocities were found between 5.54 and 5.65 km/sec. Then it was assumed that the velocity is the same everywhere in the region under consideration, equal to the weighted mean (5.574 km/sec) of the velocities found for the various groups. With this velocity, the origin time was determined once more for each shock. The resulting differences were between -0.5 and $+0.3$ seconds. Assuming these origin times, the velocity was redetermined for all 277 observed arrival times of $\overline{P}$. This gave 5.577 ± 0.002 km/sec. The figure following "$\pm$" is the probable error; its small value indicates that the data are very consistent. It does not mean that we know the velocity even within 0.01 km/sec, as the systematic errors, for example the effect of the local conditions at the stations, the variations in the thickness of the deeper layers, errors in the determination of the epicenters, are expected to be larger. Besides, at the longer distances the beginning of phases is read rather too late than too early, and consequently the velocities found are rather too small than too large; for this same reason, the origin times are calculated rather too early than too late.

The value of 5.577 km/sec was used afterwards in all calculations. The velocities found in 1932 were between 5.4 and 5.8 km/sec (average 5.55). In Table 12 residuals are given for $\overline{P}$ for the same shocks as in Table 10 for *Pn*. The data for Haiwee and Tinemaha indicate that there is no noticeable effect of the Sierra Nevada on the travel times of $\overline{P}$ to these

TABLE 12.—*Residuals of* $\overline{P}$ *in selected shocks (seconds)*

Station	Shock number (Table 3)										
	3	6	16	17	22	27	28	32	36	46	50
Pasadena	+1.0?	−0.2	−0.3	−0.3	−0.2	−0.1	−0.2	−0.2	+0.4	+0.5	+0.9
Mt. Wilson	+0.5	−0.2	−0.1	−0.2	+0.1	+0.2	−0.1	−0.2	−0.2	−0.2	+0.3
Riverside	+0.1	.....	−0.3	+0.2	+0.2	0.0	.....	−0.5	+0.2	+0.3	.....
Santa Barbara	−0.1	+0.5	+0.2	+0.3	+0.3	−0.1	−0.7	+0.3	.....	−0.1	+0.6
Haiwee	−0.1	−0.3	+0.1	+0.7	0.0	−0.2	0.0	+0.4	−0.3	−0.5	−0.1
Tinemaha	−0.3	+0.6	+0.3	+0.3	−1.0	0.0	+1.2	−0.5	0.0	−0.4	−0.4
Boulder City	−0.1	.....	+0.2	−0.1	.....	.....	.....	.....	−0.4	.....	0.0

stations. Data for stations having only a few readings for $\overline{P}$ have been omitted. (For seismograms showing *P*-phases of some of these shocks, see Gutenberg, 1943, Pl. 1.)

OTHER *P*-PHASES

In his previous paper (Gutenberg, 1932a) the author had considered three refracted longitudinal waves having their deepest points in three intermediate layers; the following were their travel times:

$$Px \qquad t = 4.9 + \Delta/7.60$$
$$Pm \qquad t = 3.9 + \Delta/6.83$$
$$Py \qquad t = 1.3 + \Delta/6.05$$

The preliminary results of the present investigation indicate that, in general, *Py* is the most outstanding of these phases both as to the amplitudes and the range over which it can be observed. (*See* Gutenberg, 1943, Pl. 1, where this wave is denoted by P^*.) The travel times of the observed *Py*-waves were investigated in groups by the method of least squares in the same way as those of the other two *P*-phases. The velocities found are between 5.92 and 6.14 km/sec with no definite indications of regional variations. All 189 data available combined in a least square solution for v and τ give 6.053 $\pm$ 0.005 km/sec. The corresponding intercept time τ is 1.29 $\pm$ 0.05 seconds; the average for shocks numbers 1 to 7 is nearly the same as for the rest.

Differences in the thickness of the granitic layer cannot be determined as accurately as differences in the depth of the Mohorovičić layer. One reason is that *Pn* is the first wave in the seismogram, and therefore can

be measured more accurately than *Py* which is superimposed on motion from earlier phases. On the other hand, an increase in the depth of the Mohorovičić discontinuity adds to the travel time of *Pn* about twice as much as an increase in the depth of the granitic layer adds to the time of *Py*. No definite conclusion can be drawn yet as to the question how much the thickness of the granitic layer varies in the area covered by this investigation. The results show only that its average thickness is about 18 km, and that local deviations from this value probably do not exceed 10 km.

The phases *Pm* and *Px* and some impulses which are suspected of being reflected longitudinal waves, have not yet been studied in detail.

The travel times of the longitudinal waves (excluding shocks and stations near the Sierra Nevada and the Little San Bernardino Mountains) indicate that in the coastal areas of southern California and in Imperial Valley the depth of the Mohorovičić discontinuity is about 40 km.

TRANSVERSE WAVES

The plotting of the transverse waves presents a rather confusing picture, which can be disentangled only in detailed research. In the case of $\overline{S}$, difficulties arise partly from possible confusion with other phases, especially surface waves, partly from the possibility that the $\overline{S}$-wave first recorded at a given station may have originated at a different place along the active trace of the fault than the first *P*-waves. In a given material faulting cannot proceed faster than with the speed of the fastest, that is longitudinal, waves; the actual speed of faulting is probably between the velocity for longitudinal and that for transverse waves. As Benioff (1938) has pointed out, thus it may happen that the *S*-waves from the end point of faulting arrive earlier at a station than those from the starting point. (For additional discussion, *see* Gutenberg, 1941, p. 263-265.) However, all longitudinal waves (*Pn*, P^*, $\overline{P}$) which are the first to arrive at any point start theoretically from one and the same point, namely the point where the fracturing begins during the earthquake and which we call "origin," "source," "focus," or "hypocenter." On the other hand, no known theory supports the assumption that longitudinal and transverse waves start simultaneously from two different points. It is extremely improbable indeed that the breaks start at two points many kilometers apart within one second, one of them without producing recorded longitudinal waves; if such a result is found from observations it is more likely that phases have been incorrectly identified, or the focal depth has been incorrectly determined, or that the instruments were not sensitive enough to record the first beginning.

The velocity for $\overline{S}$ found previously for southern California was 3.23 km/sec. For a large majority of the shocks in Table 3, this velocity has been calculated between 3.2 and 3.3 km/sec, confirming the earlier result. In most shocks the origin time calculated from $\overline{S}$ agrees with that found from $\overline{P}$ within limits of error of about ½ second. In nine shocks, $\overline{S}$ seems to have arrived systematically too early by an amount not exceeding 1 second. In some of these instances the explanation given by Benioff may explain the result, since usually most of the stations are in a limited range of azimuths; this explanation may also be correct for a number of cases where only a few stations show too early an $\overline{S}$. Further, misinterpretations may play a role; at epicentral distances near 100 km the strong phase *Sy*, indicated by "*G*" in the earlier paper, precedes $\overline{S}$; at greater distances, other *S*-phases may be mistaken for $\overline{S}$. Finally, once in a while the surface shear wave *Q* which theoretically precedes $\overline{S}$ at short epicentral distances may be unusually large and may be taken for $\overline{S}$. As a whole, there is no indication that corresponding $\overline{P}$ and $\overline{S}$ waves start at different times or from different points.

The phase *Sn*, corresponding to *Pn*, is small, and, frequently, its beginning is obscured by earlier waves; its study will require much detailed investigation. *Sy*, the phase marked previously by *Sm* (velocity about 3⅔ km/sec), is usually the most conspicuous of the *S*-group at distances between about 100 and 500 km. Detailed discussion of this and other *S*-phases must wait until further study is completed.

SUMMARY

Seismograms of 50 shocks in southern California have been investigated. The methods used are described, the elements of the shocks are given, the high accuracy of the readings is shown, and the following conclusions are drawn:

(1) Most larger shocks have occurred in a few small areas near major faults. In each group, successive epicenters are distributed over subparallel branches as well as along the major faults.

(2) Direction of fault slipping agrees with previous findings by the author; improved location of epicenters has reduced discrepancies. In all of seven shocks in northern Owens Valley (Long Valley) the western block has moved relatively northward.

(3) Travel times and velocities correspond closely to those found previously for southern California. Most shocks have originated near the bottom of the "granitic layer" at a depth of about 18 km. Local differences in travel times are investigated; the most prominent are due

either to a relatively great thickness of the surface layers above the Mohorovičić discontinuity, especially under the Sierra Nevada, or to layers (usually sedimentary) covering the granitic layer.

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STUDIES ON CARBONIFEROUS INSECTS FROM COMMENTRY, FRANCE; PART I. INTRODUCTION AND FAMILIES PROTAGRIIDAE, MEGANEURIDAE, AND CAMPYLOPTERIDAE

BY F. M. CARPENTER

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ABSTRACT

The Carboniferous insects of the Commentry Basin, France, have not been adequately described. This is the first of a series of papers in which the insect fauna is redescribed; the descriptions are based on examination of the collection in

the Muséum d'Histoire Naturelle. Part I deals with three families: (1) the Protagriidae, here removed from the Protodonata and placed in the order Palaeodictyoptera; (2) the Meganeuridae, regarded as typical Protodonata; and (3) the Campylopteridae, tentatively assigned to the order Megasecoptera. The wing and body structure of the members of these families is described and discussed in detail.

INTRODUCTION

PURPOSE OF STUDY AND ACKNOWLEDGMENTS

The fossiliferous shales at Commentry, France, have long been considered the most important formation yielding Carboniferous insects. Because of the excellent preservation of the specimens and variety of groups represented, the Commentry fauna has contributed more to our knowledge of Carboniferous insects than all other faunas combined. In spite of this and the concentration of the bulk of the material in one institution (Muséum National d'Histoire Naturelle), the literature is confused and contradictory. While studying Paleozoic insects of North America, I have frequently attempted to make comparisons with the Commentry fossils, but because of the state of the literature, the comparisons have usually been uncertain and inconclusive. To clarify the structure of these fossils I studied the Commentry insects (exclusive of roaches) at the Muséum National d'Histoire Naturelle in Paris in the summer of 1938, where I made drawings and took photographs of nearly all specimens.

This investigation was financed by grants from the Museum of Comparative Zoology and the Penrose Fund of the American Philosophical Society. I am greatly indebted to Dr. C. Arambourg, director of the Laboratoire de Palaeontologie, for permission to study the Commentry insects in the Muséum National and for placing the facilities of the Laboratoire at my disposal; and to Dr. J. Cottreau and his assistants for aid in handling the collection. Thanks are due the Agfa agency in Paris for their courtesy in providing photographic equipment and illuminators for examining and photographing the specimens.

COMMENTRY SHALES

The successful collection of insects in the Commentry shales has been largely due to the fact that for many years the coal associated with the shales was mined in open pits. Most of the insects from Commentry (about 1500 specimens, including roaches) were obtained from the pits. The coal basin was not extensive, and even by 1909 (Stevenson, 1910) most of the quarries had been abandoned and utilized as receptacles of waste from other mines and industrial plants in Commentry. To my knowledge no insects have been found in the shales for the past 28 years, and according to information obtained by Doctor Arambourg all mining

at the basin has been discontinued. Probably no further Commentry insects will be found.

Geologists have differed as to the history of the Commentry basin (Stevenson, 1910) but have agreed that it was originally the site of a deep, fresh-water lake, about 5½ miles long, almost 2 miles wide, and partially surrounded by mountains. Several streams entered the lake, which, according to Stevenson, was rimmed by a low beach backed by a marsh. The plants consisted of Cordaites trees, ferns, lepidodendra, and other vegetation.

The Commentry shales have been referred consistently to the Stephanian series, which is generally assigned to the upper part of the Carboniferous (Pennsylvanian).[1]

PREVIOUS WORK

Our knowledge of the Commentry insects began with the studies of Charles Brongniart, who described 97 species (other than roaches) between 1877 and 1894. Most of these were first mentioned very briefly in a series of short publications before 1886, but all were redescribed more fully in his elaborately printed *Recherches* in 1894.[2] In spite of its excellence this work has certain defects. First, the descriptions are inadequate for our present needs; there is no mention of the convexity or concavity of the veins, though these are sometimes indicated by shading in the drawings, or of the state of preservation—for example, whether or not the origins of certain veins are absolutely clear in a specimen. Second, though Brongniart obviously depended upon illustrations for depicting details, many figures are unsatisfactory. (1) With few exceptions, the drawings are natural size, a distinct disadvantage for specimens less than 2 or 3 cm. long, as details cannot be ascertained. (2) Although most wing drawings are accurate, some are incorrect, for example, those of *Meganeura monyi* on Plate XLI (25), a discussion of which is given below. (3) The artist who made drawings of the fossils *in situ* apparently took undue liberty in depicting the position of the speci-

[1] Handlirsch, though originally following this concept, subsequently (1920; 1922) placed the Stephanian series in the Middle Carboniferous, contemporaneous with the late Westphalian. As Lameere (1917) pointed out, however, nothing in the Commentry fauna justifies this view.

[2] The publication date has consistently been cited as 1893, given on the title page, but conclusive evidence at hand shows that it really was not published until late 1894. Page 12 bears a reprint of an article by Scudder from the American Journal of Science for February 1894. Furthermore, a proof of Plate 23, autographed by Brongniart and now in my possession, is dated April 2, 1894, by the photoengraver Dujardin. Brongniart's letters to Scudder, placed at my disposal by the New England Museum of Natural History, set the time of publication even later than this. In a letter dated May 10, 1894, Brongniart wrote Scudder as follows: "Mon ouvrage sur les insectes primaires est imprimé . . . je pense que mon ouvrage paraitra dans les premiers jours de juin; je vous en enverrai immédiatement un exemplaire." Finally, in a letter dated January 30, 1895, Brongniart wrote that his work would reach Scudder in the course of February, and Scudder (1895) recorded the receipt of the work in June 1895. The exact date of publication is still not certain, but it was clearly not before June 1894.

mens on the rock. (*See* specimens of *Meganeura monyi*.) This freedom of illustrating has led to serious errors; thus, in some cases a wing preserved in pieces or on separate fragments of rock which do not fit at all closely together has been drawn as a continuous wing on an unbroken slab. This has caused confusion and misunderstanding.

Following the publication of Brongniart's *Recherches*, nothing was added to the literature on the Commentry fauna until Agnus (1902; 1904) described one palaeodictyopteron (*Homoioptera gigantea*) and a few roaches in the collection of the Muséum National (Paris). In 1907 Fernand Meunier began publication on the extensive series of specimens received at the Muséum the preceding decade. During the next 7 years he described 39 new species (apart from roaches). Most of his descriptions, like Brongniart's, first appeared in short articles, but all were republished in his *Nouvelles Recherches* (1909; 1912), together with superb lithographs of nearly all specimens. Meunier's examination of the fossils, however, was obviously superficial (Handlirsch, 1906b), and his descriptions, though longer than Brongniart's, usually omit the important characteristics. His drawings are diagrammatic and unreliable. The description of *Archaeoptilus gaullei* Meunier (1912, p. 233) is a good example. This species was actually based on the counterpart of the type of Agnus' *Homioptera gigantea*, which was also in the Paris Muséum. A comparison of Meunier's drawing of this fossil with the published photographs (Meunier, 1912, Pl. 6, fig. 1; Agnus, 1902, Pl. 1) reveals the defects of the former; even the prothoracic lobe has been completely omitted. The taxonomic confusion of the Caloneuridae typifies the disorder resulting from Meunier's superficial work on the Commentry insects (Carpenter, 1934, p. 325-326).

No insects have been collected in the Commentry shales, to my knowledge, since 1914; hence, subsequent publications have dealt with previous collections and, in regard to the material in the Paris Muséum, with the specimens described by Brongniart and Meunier.[8] During part of the war (1916), Professor August Lameere, of Brussels, took advantage of a forced sojourn in Paris to study all the Commentry insects (except the roaches) in the Muséum National, chiefly because of the diverse conclusions reached by Brongniart and Handlirsch on the classification of the Commentry fossils. The results are contained in his *Revision sommaire des insectes fossiles du Stéphanien de Commentry* (1917), which, although only 60 pages long and without figures, has been most useful. In addition to descriptions of nine new species,

[8] Part 3 of Meunier's *Nouvelles Recherches*, though not published until 1921, was actually completed in 1914.

it contains notes on nearly all Brongniart's and Meunier's types, with some indications of synonymy. Lameere's observations are accurate and agree with mine in most cases. However, because of the absence of illustrations and the brevity of his descriptions, this summary does not give a working knowledge of the Commentry fauna.

The literature previously discussed has dealt with the Commentry insects in the Muséum National d'Histoire Naturelle. A few additional specimens are in the British Museum and the Manchester Museum —17 in the former and 9 in the latter. Five species (apart from roaches) were described by Bolton (1917; 1925) from these two collections, and another was added by Handlirsch (1919) from the British Museum material.[4] Two species described by Bolton in the Manchester Museum were subsequently redescribed by Tillyard (1918).

The Commentry insects (totalling 152) described in the publications mentioned were based on examination of the actual fossils. In addition, 37 species were erected by Handlirsch (1906; 1919) for specimens which he did not see but which were figured or mentioned by Brongniart or by Meunier and identified by them as conspecific with other Commentry insects. This procedure, which Handlirsch has unfortunately followed in many of his works on fossil insects, has been the main reason for the chaotic state of the literature on the Commentry fauna. In his *Revision de Paläozoischen Insekten* (1919), Handlirsch included drawings of most of the 35 Commentry species therein established, but, although it is not so stated, all were based on the photographs and other figures published in the works of Brongniart and Meunier. Handlirsch's attempt to describe the species without seeing the specimens would be ludicrous if it had not been so confusing. One example will demonstrate this. Although *Meganeura monyi* was originally established by Brongniart on six fragments of wings, Handlirsch (1906; 1919) restricted *monyi* to one of these specimens and established a new species for each of the other five specimens of the original type series. However, three of these six fossils were really counterparts of the other three. Handlirsch therefore actually established six species for three specimens, describing three of the fossils as species distinct from their counterparts. Many of Handlirsch's genera and even families were erected for Commentry fossils with almost as little basis as these "species."

As already indicated, 189 species of insects (exclusive of roaches) have been described from the Commentry shales. Their classification

[4] The type of Handlirsch's species *(Microdictya latipennis)* (1919) was identified as *Goldenbergia vaillanti* Brongniart by Bolton (1925), who was not aware of Handlirsch's description of the fossil.

into genera and especially families is very difficult; not enough is known about most of the species or their relatives from other Carboniferous strata to enable us to determine their affinities with certainty. However, with the knowledge of Permian insects acquired in the past 15 years, we are better able to arrive at a satisfactory classification than Brongniart and Lameere were in 1894 and 1917. The classification which Lameere adopted after his study of the Commentry fossils (1917) differs widely from that proposed for Carboniferous insects in general by Handlirsch, Tillyard, Martynov, and others. I hope that the present survey of the Commentry fauna, with the aid of our newly acquired knowledge of the Permian insects, will clarify some of the obscure and doubtful aspects of the classification of these insects and those of other Carboniferous strata.

FAMILIES PROTAGRIIDAE, MEGANEURIDAE, AND CAMPYLOPTERIDAE

GENERAL STATEMENT

Among the Commentry insects described by Brongniart four genera, *Protagrion, Meganeura, Titanophasma,* and *Campyloptera,* were placed in the "family" Protodonata. The latter group has subsequently been raised to ordinal rank (Handlirsch, 1906), and the families Protagriidae, Meganeuridae, and Campylopteridae erected for these and related genera. Although opinions differ as to the systematic position of these families, it has been generally agreed that they are more or less closely related. Examination of the type specimens on which the genera were based has led me to alter the views I have previously held with other students of fossil insects (Carpenter, 1931; 1939). I now believe that the family Protagriidae should be moved from the Protodonata to the order Palaeodictyoptera, and that the Meganeuridae should be regarded as typical Protodonata. I am still uncertain of the ordinal affinities of *Campyloptera.* A new order will probably be needed eventually for this remarkable insect, but, until we know more about its structure, I prefer to place it among the Megasecoptera. Reasons for these conclusions will be discussed below.

FAMILY PROTAGRIIDAE (ORDER PALAEODICTYOPTERA)

Systematic descriptions.—This family is monotypic, including only *Protagrion audouini* Brongniart from Commentry.[5] Since the structure of this insect has not

[5] The Permian genera *Calvertiella* Tillyard and *Tillyardiella* Martynov were at one time referred to this family also, but both of these have subsequently been placed in separate families (Martynov, 1931; Handlirsch, 1937). *Tillyardiella* is so incompletely known that even its ordinal affinities are uncertain.

been correctly understood in the past, the following diagnosis is suggested for the family:

Allied to Dictyoptilidae.

WINGS: precostal space absent; Sc terminating near the apex of the wing; stems of R and M coalesced; Rs with numerous branches; MA simple, anastomosed with Rs near the origin of the latter; MP with numerous branches; CuA simple, arched anteriorly shortly after its origin and almost touching R + M; CuP branched; 1A without true branches; cross-veins numerous. Only one pair of wings, apparently the fore wings, are known, but the other pair probably had a similar venation.

Genus *Protagrion* Brongniart

Protagrion BRONGNIART, 1894, Recherches Hist. Ins. Foss., p. 403; HANDLIRSCH, 1906, Foss. Ins., p. 306.

WINGS: Slender; Rs with about nine branches, MP with about seven; anastomosis of Rs with MA very short; CuA connected to R + M by a short cross-vein just basad to separation of M from R; cross-veins forming a coarse reticulation near apex of wing.

Protagrion audouini Brongniart

(Figure 1)

Protagrion audouini BRONGNIART, 1894, Recherches Hist. Ins. Foss., p. 403; pl. 24, figs. 1, 2; HANDLIRSCH, 1906, Foss. Ins., p. 306; pl. 31, fig. 36; LAMEERE, 1917, Mus. Nat. Hist. Natur., Bull., vol. 23, p. 162; HANDLIRSCH, 1919, Denkschr. Akad. Wiss., vol. 96, p. 57; TILLYARD, 1925, Am. Jour. Sci., 5th ser., vol. 10, p. 47, fig. 3; TILLYARD, 1928, Rec. Ind. Mus., vol. 30, p. 159, fig. 3; TILLYARD, *in* TILLYARD AND FRASER, 1938, Austral. Zool., vol. 9, p. 130, fig. 3.

This species is known only from the unique type (obverse and reverse) in the Muséum National at Paris. Parts of two wings, probably the fore pair, and one leg are preserved. One wing, 90 mm. long, is almost complete, lacking only portions of the posterior margin. Figure 1 includes a drawing based on both wings of the fossil. Brongniart's figure of *audouini* is crude and does not give accurate details of venation. I believe the wing was more slender than he depicted it. Most of the posterior margin is unknown, except for the apical fifth of the wing and a small piece near the middle. Brongniart probably mistook for the hind margin a slight irregularity in the rock matrix, since the veins do not extend to the level of the margin, as he draws it, but only to the faint line which I have interpreted as the margin. On this basis, the wing is 18 mm. wide instead of 21 mm., as he thought.

In his figure of *audouini* Brongniart showed a fairly large precostal area on the left wing and a smaller one on the right. I believe, however, that no precostal area existed in this species; in spite of careful search, I was unable to find even a trace in either wing. The proximal parts of both wings are well preserved, and I cannot understand what led Brongniart to draw these areas. The fact that they are of different sizes and shapes in his drawing suggests that he mistook some peculiarity of the matrix for part of the wing.[6]

Sc cannot be traced to its termination in either wing but can be followed very nearly to the apex, where it almost certainly terminates on the costal margin. Rs gives rise to nine terminal branches in that wing which has the apex preserved, but the number of branches probably varied in individuals; MA is anastomosed with Rs

[6] The Permian *Calvertiella*, generally considered the closest known relative of Protagrion, also lacks a precostal area.

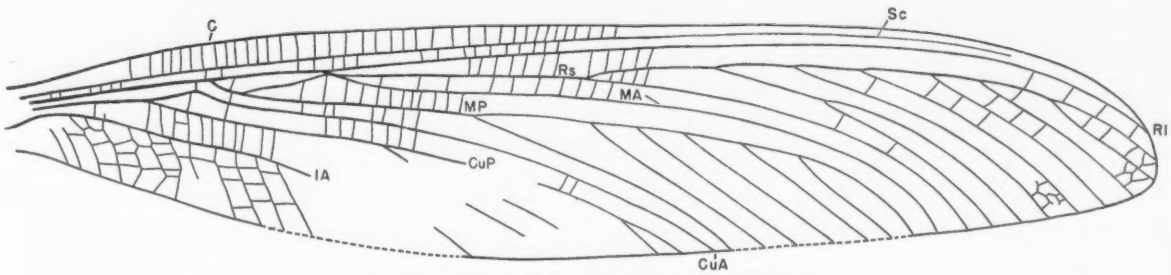

FIGURE 1.—*Wing of* Protagrion audouini *Brongniart*

Original figure, drawn from type. (C) costa; (Sc) subcosta (—); (R1) radius (+); (Rs) radial sector (—); (MA) anterior media (+); (MP) posterior media (—); (CuA) anterior cubitus (+); (CuP) posterior cubitus (—); (1A) first anal (+).

for a very short distance, almost at the latter's origin; CuA is directed anteriorly soon after its origin and almost touches R+M. 1A is connected to the posterior margin by a series of veinlets forming with the cross-veins a coarse network of irregular cells.

Although Brongniart's figures of *audouini* give the general features of the venation correctly, they are too small to represent the details accurately. Subsequent workers who have depended on these drawings for their knowledge of the venation have been misled. The figures given by Handlirsch (1906; 1919) are based on Brongniart's but the origins of M, CuA, and CuP and the shape of the posterior margin of the wing are incorrectly shown. Tillyard's figures of the wing (1925; 1928; Tillyard and Fraser, 1938) are much more incorrect than Handlirsch's. Not only are the latter's errors repeated, but others, such as a simple CuP, are added.[7] Lameere's brief description of *audouini* (1917) agrees fully with my observations; as he points out, not only does MA diverge anteriorly and coalesce with Rs basally, but CuA similarly diverges and is connected to R+M by a very short cross-vein. Another important characteristic mentioned by Lameere is the branching of CuP (Cu2). This vein had been represented in all Tillyard's figures as unbranched (as it is in the Meganeuridae and Odonata), but it is definitely branched several times in the fossil. The beginning of one of the branches is clearly visible in one wing, and the distal contour of the preserved part of 1A shows that this vein must terminate far basad to the distal end of CuA. This branched CuP has much phylogenetic significance, as shown later.

Affinities of Protagriidae.—Brongniart believed that *Protagrion* was closely related to the genus *Paralogus* Scudder from the Carboniferous of Rhode Island. The latter, however, lacks MP and CuA (which are present in *Protagrion*) and in recent years has consequently been placed in the family Meganeuridae. *Protagrion* was separated by family rank from *Meganeura* and its relatives in 1906, and, after Lameere's statement of the presence of MP and CuA in *Protagrion* (1917), it was further separated from the Meganeuridae. Martynov in 1932 removed the Meganeuridae from the Protodonata and established for them the order Meganisoptera, leaving the genus *Protagrion* and the recently described *Calvertiella* in the order Protodonata. Tillyard (Tillyard and Fraser, 1938) proposed that the Meganisoptera be considered a suborder of the Odonata instead of a separate order. About the same time (1938) I independently suggested the division of the order Protodonata into two suborders: Euprotodonata, to include the Protagriidae and Calvertiellidae, and Meganisoptera, for Meganeuridae and its relatives. However, study of the specimen on which *Protagrion* is based has convinced me that it is not so closely related to the Meganeuridae or Odonata as supposed and that it is really a member of the order Palaeodictyoptera, its nearest relatives being the Dictyoptilidae. A similar association of *Protagrion* with *Dictyoptilus* was implied in Lameere's summary of the Commentry insects in 1917, for, although he eliminated the order Palaeodictyoptera from his classification, he grouped *Protagrion* and *Meganeura* into the family Dictyoptilidae, which also included three genera (*Dictyoptilus, Peromaptera,* and *Archaemegoptilus*) usually placed in the Palaeodictyoptera. The family Dictyoptilidae he regarded as identical with the Protodonata. In a later paper (1935), however, he recognized a group termed the Odonatoptères, which he divided into (1) Stenodictyoides, including

[7] Tillyard stated (1928, p. 159) that his drawing was based on a study of the fossil and a photograph of it, but the striking discrepancies between his figures and Brongniart's as well as the observations made by Lameere and me indicate that his examination of the specimens was not careful. To my knowledge this specimen of *audouini* is the only Commentry fossil in the Muséum National which he reported having examined.

Dictyoptilus, and so forth, and (2) Odonatoides, consisting of the Odonata and Protodonata, including *Protagrion* and *Meganeura.* This has been the conventional view of the relationships of *Protagrion.*

Three venational features of *Protagrion* have led me to place the genus in the Palaeodictyoptera instead of in the Protodonata. (1) Presence of a branched CuP. Nowhere among the true Odonata or Meganeuridae is a branched CuP known, the simple form of this vein being a major characteristic of the odonate complex (Meganeuridae and Odonata). A branched CuP does occur, however, in *Dictyoptilus, Microdictya,* and related Commentry genera which have been regarded as typical Palaeodictyoptera. (2) Proximal structure of MA and CuA. Both diverge anteriorly shortly after their origins, and MA anastomoses with Rs. Nothing of this nature has been observed anywhere in the Odonata or Meganeuridae.[8] A comparison of the venation of *Protagrion* with that of *Dictyoptilus,* however, shows striking similarities in the formation of these veins. MA in *Dictyoptilus* is almost in contact with Rs; and further coalescence of M with R (a coalescence already started in this genus) would result in a configuration almost identical with that in *Protagrion.* (3) Presence of MP and CuA in *Protagrion.* These veins are lacking in the Odonata as well as in the Meganeuridae, a condition which has already led to the separation of *Protagrion* from the Meganeuridae by subordinal or even ordinal rank. It is especially noteworthy, however, that MP is not a simple vein in *Protagrion* but is well developed, with fully seven branches, and gives no indication of approaching an obsolescent condition. Both MP and CuA are present in *Dictyoptilus,* however, and the former is a well-developed vein.

The wings of *Dictyoptilus* and *Protagrion* have other features in common, such as the narrowed costal space at the base, the arched branches of Rs, a simple MA, and even a similar arrangement of cross-veins.

The family Calvertiellidae, with the single genus *Calvertiella,* should also be considered palaeodictyopterous until evidence to the contrary has been found. Handlirsch (1937) has already placed the family in that order. On the other hand, I believe the affinities between *Protagrion* and *Calvertiella* are not nearly so close as Tillyard supposed. *Calvertiella* differs from *Protagrion* in several important respects: Sc terminates on R1 not far beyond midwing, intercalary sectors are present between the branches of Rs, CuP is unbranched, and the whole pattern of the crossveins is very different.

Since the order Protodonata was originally established for *Meganeura* as well as *Protagrion,* the transfer of the latter to the Palaeodictyoptera does not place the order Protodonata in synonymy. Martynov's order Meganisoptera, however, becomes synonymous with the Protodonata, whether the latter be regarded as an order or as a suborder of the Odonata.

FAMILY MEGANEURIDAE (ORDER PROTODONATA)

Systematic descriptions.—This family, although originally established by Handlirsch for a few species in the Commentry shales, now includes many genera from the Permian and Carboniferous of Europe and North America. Six genera, including 12 species, have been recognized in the Commentry fauna, but I am convinced

[8] In his summary of the Commentry insects in 1917, Lameere suggested that the anastomosis of MA and Rs in *Protagrion* might represent an early stage of the evolutionary sequence leading to the crossing of MA by a branch of Rs in the Odonata. In more recent years, however, this supposed crossing in the Odonata, originally proposed by Comstock and Needham, has been disproved. (*See* Carpenter, 1931.)

for reasons given below that only two genera and four species are valid. These two genera, *Meganeura* and *Meganeurula,* differ from each other chiefly in the structure of the precostal area and the origin of R4+5; *Meganeura* is consequently placed in the subfamily Meganeurinae, and *Meganeurula* in the subfamily Typinae.

Subfamily MEGANEURINAE
Genus *Meganeura* Brongniart

Meganeura BRONGNIART, 1885, Soc. Amis Sci. Nat. Rouen, Bull., vol. 21, p. 60; BRONGNIART, 1894, Recherches Hist. Ins. Foss., p. 396; HANDLIRSCH, 1906, Foss. Ins., p. 307; LAMEERE, 1917, Mus. Nat. Hist. Natur., Bull., vol. 23, p. 161.
Meganeurella HANDLIRSCH, 1919, Denkschr. Akad. Wiss., vol. 96, p. 59.

FORE WING: Slender; precostal space very broad at base, probably extending to midwing or slightly beyond; Sc remote from R+M proximally, long, and extending at least almost to apex of wing; Rs and MA arising from R+M at about the same point; R4+5 arising very shortly after the origin of Rs, well before the level of the origin of the first branch of MA; R3 originating well beyond the middle of the wing; proximal part of the anal area sclerotized, forming a distinct, veinless area. Crossveins very numerous.

HIND WING: Broader than the fore, especially proximally; precostal area narrower than in fore wing, extending to about midwing; origins of Rs, MA, R4+5, and R3 essentially as in fore wing; anal area nearly three times as wide as in the fore wing.

The body structure is very incompletely known, but it was probably not unlike that of existing Odonata. The mandibles were certainly well developed; the eyes large, nearly contiguous. The antennae are unknown. The prothorax was shorter than the mesothorax or metathorax. The thorax as a whole was apparently as oblique as that of the anisopterous Odonata, the legs attached far anteriorly; all legs were armed with strong spines, but the tarsal segmentation is unknown. The abdomen is not preserved in any specimens.[9]

Brongniart originally erected this genus for a single species, *monyi,* from the Commentry shales, though he later added a second species, *selysii.*[10] His definition of *Meganeura* was very broad, agreeing almost perfectly with our present concept of the whole family Meganeuridae. Handlirsch, in 1906, redefined *Meganeura* and established a new genus (*Meganeurula*) for *selysii,* but, not having seen the fossils, he based his concept of the genera on Brongniart's drawings. The latter, unfortunately, were not correct in several important respects, as mentioned in part by Lameere, so that Handlirsch's account of *Meganeura* does not entirely agree with the genotype specimens. Tillyard, who also published a general account of *Meganeura,* was similarly misled by Brongniart's drawings. This confused state of the literature became chaotic about a decade later when Handlirsch (1919) erected three additional Commentry genera (*Meganeurites, Meganeurina,* and *Meganeurella*), basing them on Meunier's published photographs and crude drawings. After examining the fossils, all of which are in the Muséum National d'Histoire Naturelle, I am convinced that only one of Handlirsch's genera (*Meganeurula*) is valid.

[9] The isolated body of a large Commentry insect, *Titanophasma fayoli* Brongniart, although originally placed by Brongniart among the orthopteroids, was later (1894) considered by him to be a meganeurid. This view was shared by Lameere (1917), who believed it was almost certainly the body of a species figured by Meunier (1909, Pl. 4, fig. 1). However, I believe not enough is preserved in the type of *T. fayoli* to indicate even its ordinal position.

[10] Brongniart first (1885) considered *Meganeura* to be orthopteroid (Protophasmida) but later (1894) recognized its true affinities.

Before presenting the evidence for this conviction and discussing the characteristics of *Meganeura*, I wish to call attention to two facts which have an important bearing on the conception of the genus. One deals with Brongniart's drawings of the types; the other concerns the fossils themselves.

(1) Brongniart (1894, Pl. 25, figs. 1-6) includes drawings of the six specimens comprising the type series of *monyi*. In his text he gives no account of the nature of the individual specimens—*i.e.*, condition of preservation—nor does he explain any of the obvious differences between his figures of the wing fragments, apparently assuming that the illustrations were sufficiently informative. I was therefore much surprised to find that the pieces of rock bearing the fossils are not at all like the drawings on Brongniart's plate. Thus, Figure 3 shows three parts of a wing on one piece of rock; actually, however, these fragments are on three separate pieces of rock, undoubtedly broken apart when collected. The same is true of the fossil shown in his Figure 2. This discrepancy between the drawings and the actual fossils must be borne in mind whenever reference is made to Brongniart's figures. It explains certain differences between the individual wings as they are represented by him. For example, the proximal piece of the wing in Figure 2 is drawn much too close to the middle piece. The wing as figured therefore has very different proportions from those of other wings. Brongniart's drawing, however, shows the proximal wing fragment on the same continuous piece of rock as the proximal piece; if this were actually so, there could be no question of the distance between the two wing fragments or of the shape of the wing. One might ask if these wing fragments were really not broken apart after the drawings had been made. This would be a logical explanation, if it were not for the fact that every piece of rock depicted on Plate 25 has a very different shape from that shown in the drawing. In all instances, the wing itself extends to the very edge of the rock, at least on one side, whereas the drawings show the wings bordered by considerable matrix.

(2) When referring to Brongniart's figures of *monyi*, I have often wondered if some of the type fossils might not be counterparts of others in the series; this seemed to be entirely precluded, however, by Brongniart's drawings. On noting the discrepancy in shape between the fossils and Brongniart's drawings, however, I attempted to fit certain of the specimens together and found that three of them (shown in Brongniart's Figs. 4, 2, 5) actually were counterparts of the other three (Figs. 1, 3, 6, respectively). Hence the type series of *monyi* really consists of three individual specimens instead of six as previously supposed. This has much taxonomic significance, inasmuch as each of these six has been made the type of a distinct species by Handlirsch.

Returning to Handlirsch's four genera mentioned above, we can readily understand that, having seen neither the specimens of *monyi* nor the fossils for which he erected the genera, Handlirsch had little factual knowledge of the species concerned. Brongniart's drawings of *monyi* were incorrect in several respects, the most important being the structure of R1 in the fore wing. This vein is branched in his Figures 1 and 3 (Pl. 25), though it is correctly shown as unbranched in his restoration on the following plate. Handlirsch, assuming the restoration to be wrong, characterized *Meganeura* as having a branched R1 in the fore wing and a simple one in the hind wing and erected the genus *Meganeurula* for *M. selysii*, since it had a simple R1 in both pairs of wings. However, as Lameere has pointed out, the structure of R1 in *monyi* is exactly like that of *selysii*. The latter does differ from *monyi*, however, in having a small precostal area. This characteristic, overlooked by Brongniart, Handlirsch, and Lameere, is sufficient basis for the retention of the genus *Meganeurula*.

Meganeura monyi (Brongniart)

(Figure 2)

Dictyoneura monyi BRONGNIART, 1884, Acad. Sci., C. R., vol. 98; p. 833.
Meganeura monyi BRONGNIART, 1885, Soc. Amis Sci. Nat. Rouen, Bull., vol. 21, p. 60; BRONGNIART, 1894, Recherches Hist. Ins. Foss., p. 161, pl. 25, figs. 1-6; pls. 26, 27; HANDLIRSCH, 1906, Foss. Ins., p. 307; pl. 32, fig. 2; MEUNIER, 1909, Ann. Paleont.,

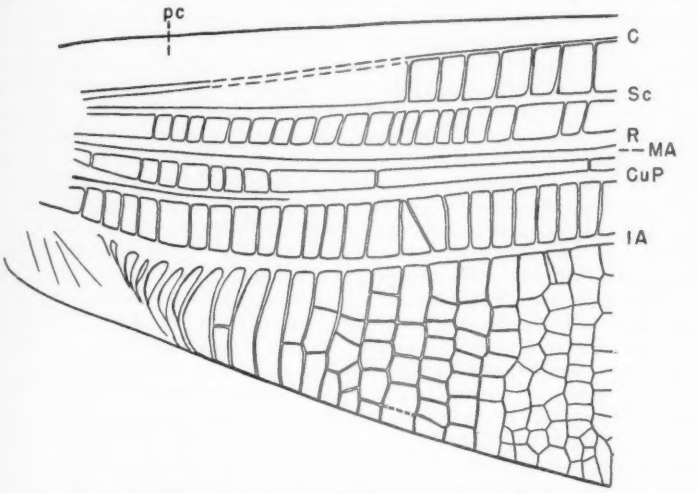

FIGURE 2.—*Base of fore wing of* Meganeura monyi *Brongniart*
Original figure drawn from Meunier's specimen. (pc) precostal space; other lettering as in Figure 1.

vol. 4, p. 144; pl. 3, fig. 2; LAMEERE, 1917, Mus. Nat. Hist. Natur., Bull., vol. 23, p. 161; TILLYARD, 1917, Biol. Dragon-flies, p. 304, fig. 156; HANDLIRSCH, 1919, Denkschr. Acad. Wiss. Wien, vol. 96, p. 58, fig. 61; VIGNON, 1929, Mus. Hist. Nat., Arch., pl. 2, fig. 4; CARPENTER, 1938, Am. Acad. Arts Sci., Pr., vol. 73, p. 40.
Meganeura brongniarti HANDLIRSCH, 1906, Foss. Ins., p. 308, pl. 32, fig. 3; HANDLIRSCH, 1919, Denkschr. Acad. Wiss. Wien, vol. 96, p. 58, fig. 62; LAMEERE, 1917, Mus. Nat. Hist. Natur., Bull., vol. 23, p. 161.
Meganeura brongniartiana HANDLIRSCH, 1919, Denkschr. Acad. Wiss. Wien, vol. 96, p. 58, fig. 63.
Meganeura fafnir HANDLIRSCH, 1906, Foss. Ins., p. 58, pl. 32, fig. 4; LAMEERE, 1917, Mus. Nat. Hist. Natur., Bull., vol. 23, p. 161; HANDLIRSCH, 1919, Denkschr. Acad. Wiss. Wien, vol. 96, p. 58, fig. 64.
Meganeura draco HANDLIRSCH, 1919, Denkschr. Acad. Wiss. Wien, vol. 96, p. 59, fig. 65.
Meganeura aeroplana HANDLIRSCH, 1919, Denkschr. Acad. Wiss. Wien, vol. 96, p. 59, fig. 66.
Meganeurella rapax HANDLIRSCH, 1919, Denkschr. Acad. Wiss. Wien, vol. 96, p. 59, fig. 67.

This species is known only from fragments, which even when combined do not represent a complete fore or hind wing. Because of this and our unsatisfactory knowledge of related Meganeuridae, it is impossible to determine good specific characteristics of *monyi*. The size and shape of the wings are all that can be indicated

in this category. The place of origin of some of the main veins and the nature of the cross-veins probably have some specific value also, but until more species of the genus are known distinction between specific and individual traits is only tentative.

WINGS: The fore wing of *monyi* is known from two specimens; one of these is represented in Brongniart's Figures 1 and 4 of Plate 25 (1894); the other is the Meunier specimen depicted in Figure 2 of Plate 3 (1909). Neither of these shows more than about one-fourth or one-third of the wing. The hind wing is known from three Brongniart specimens: (1) Figures 1 and 4 of Plate 25; (2) Figures 2 and 3; (3) Figures 5 and 6. Although more of the hind wing is known than of the fore, our knowledge of it is far from complete; the posterior border of the wing is not preserved in any specimen except at the very base of the wing, and the apex is entirely unknown. It is therefore apparent that the shape and size of the wings of *monyi* can be determined only approximately. The largest continuous fragment of either wing is preserved in the specimen shown in Figure 1 of Plate 25; this is a hind wing, 260 mm. long. A comparison of its venation with that of more nearly completely known meganeurids indicates that the hind wing of *monyi* was about 300 mm. long. The shape of the wings is much more difficult to determine. Nearly the proximal third of the fore wing is preserved in two specimens (Brongniart's Pl. 25, fig. 1; Meunier's Pl. 3, fig. 2). These show that the fore wing was decidedly slender, with a narrow anal area. There is no evidence whatever for the abruptly widened anal area shown in Brongniart's restoration of the wing (Pl. 26), and I cannot understand why the wing was so drawn by him and by others (Handlirsch, Tillyard) who have published restorations "modified from Brongniart."[11] The fore wing as a whole was probably much more slender than assumed if the fore wing of *selysii* or other meganeurids is indicative of that of *monyi*. The apex of the wing is not known in either *selysii* or *monyi*; but in *Megatypus* and *Typus* (Permian) it is pointed, and this was probably also true of *Meganeura*.

It is therefore apparent that the published restorations of the wings of *Meganeura* (Brongniart, 1894, Pl. 26; Handlirsch, 1921, Fig. 72; Tillyard, 1917, Fig. 156) are distinctly inaccurate in several respects, as well as highly imaginative in others. The very schematic drawing of the *Meganeuridflugel* by Handlirsch (1919) represents more nearly than these the correct form of the wing (at least basally), though many of the venational features shown are not those of *monyi*.

The drawings of the types of *monyi* included on Brongniart's Plate 25 contain several serious errors which require notation to prevent further confusion.

SPECIMEN 1.[12] The piece of rock on which this specimen is preserved is much more irregular than represented in the drawings; both fore and hind wings extend to the margin of the rock, which traverses them at about a 45-degree angle. The chief inaccuracy of the drawing is the branching of R1 in both fore wings. The branch indicated really arises from Rs and is R4+5. The drawing of the reverse piece (Fig. 4) is essentially correct except that it does not depict the anal area clearly.

SPECIMEN 2. This wing is preserved in several pieces, which do not fit together as closely as they are drawn in Brongniart's figure. The proportions of the proximal piece are incorrectly represented in his figure;[13] this part of the wing is much longer

[11] A narrow anal area, like that actually present in *monyi*, is characteristic of all the meganeurid fore wings thus far known, whether Permian or Carboniferous.

[12] Since the Commentry insects in the Muséum National are not numbered or catalogued, I have used here the figure number on Plate 25 for the number of the specimen—*i. e.*, specimen 1 is the fossil represented in Figure 1 of this plate.

[13] A photograph of this specimen is reproduced by Vignon (1929, Pl. 2, fig. 4).

than drawn, and the contour of the veins is identical with that of the above specimen. Specimen 3, the counterpart of specimen 2, really consists of three wing fragments on separate pieces of rock, not on one continuous piece as shown in Brongniart's figure; and some of the wing between fragments is missing. In Brongniart's drawing the basal piece of the wing is too near the middle piece, giving wrong proportions to the wing as a whole. In his drawing of specimen 3, Rl is branched distally, though this part of the wing is not preserved at all. In his restoration of the wing Brongniart correctly showed this branch arising from Rs. Figure 3 is further incorrect in showing the precostal area extending nearly to the apex and the termination of Rl and Sc at about the same point.

SPECIMEN 5. This and its counterpart, specimen 6, are accurately drawn except for the shape of the rock containing the fragments.

In addition to the three specimens constituting the type series, *monyi* is also represented by a specimen figured by Meunier (1909, Pl. 3, fig. 2). This shows about the proximal quarter of one fore wing and somewhat more of the other, but the latter is badly distorted. The smaller part of the fore wing, however, is very well preserved and shows details which are not visible in other specimens and which deserve special comment (Fig. 2). (1) The remarkable structure of the anal area basally is clearly shown; the more proximal veinlets from 1A to the posterior margin are very thick and merge distally, forming the outer margin of what seems to be a prominent sclerotized area. This area was obviously formed by the broadening and fusion of the veinlets even nearer the base of the wing. It is apparently the homologue of the membranule occurring in the anisopterous Odonata, though it is much larger. It is clearly preserved in Brongniart's specimen 1 (4) and is crudely represented in his drawing (Fig. 4); but curiously enough it has been omitted from all other drawings and restorations of *monyi*. As already indicated above, I consider this sclerotized area a generic characteristic. Whether the base of the anal area in the hind wings of *Meganeura* was similarly formed, however, we have no means of knowing, inasmuch as the proximal part of this wing is not preserved in any specimen. I suspect that it was, for in the Anisoptera the membranule is larger in the hind wing than in the fore. (2) The proximal part of Cu in this specimen is interesting because it appears to be made up of two contiguous veins. In many meganeurid genera like *Typus* and *Megatypus* a weak but distinct vein arises at the base of the wing just above CuP and then coalesces after a short interval with the latter. This has been interpreted as the vestige of CuA. The "double" stem of CuP in *monyi*, previously unobserved in *Meganeura*, seems to present an analogous condition, except that the "vestigial CuA" is here in actual contact with CuP for its (CuA's) entire length. (3) The obliquity of some of the cross-veins between CuP and 1A is noteworthy. In most Protodonata one or two of these cross-veins, in the region just before the origin of Rs, assume an oblique position. Their significance has been extensively discussed (Carpenter, 1931, p. 101, 112), and I have shown that in *Typus* there is much variation in the position of the cross-vein. Lameere (1922, p. 10) was the first to call attention to such a cross-vein in the specimen of *monyi* figured by Meunier (1909, Pl. 3, fig. 2), here illustrated in Figure 2. It is significant, however, that in some other specimens of *monyi* (*e.g.*, Brongniart's specimens 1, 2) two oblique cross-veins are present and that their angle of inclination is not always the same. Apparently, therefore, the position of these oblique veins in *Meganeura* is as variable as that in *Typus*.

BODY STRUCTURE: Very little is known of the body structure of *monyi*. Brongniart's specimen 1 includes a vague indication of part of the thorax, but nothing definite

is preserved. His reconstruction of the body of *monyi* (Pl. 28) was obviously based in part on the type of *selysii*, which shows the head, thorax, and two pairs of legs; but for the most part it was highly imaginative. The same is true of Tillyard's restoration (1917, Fig. 132). Meunier's specimen of *monyi* (1909, Pl. 3, fig. 2) shows the prothoracic and mesothoracic legs, but the head and abdomen are entirely missing, and the thorax is so distorted as to be almost unrecognizable. What appears to be the prothorax is about 20 mm. long and 23 mm. wide. The fore and middle pairs of legs as preserved are far forward; since this is also true of the legs in the two specimens of *Meganeurula selysii*, it indicates that the thorax of these meganeurids was as oblique as in the Recent Anisoptera. The fore legs of Meunier's specimens have the same dimensions as the middle pair: the femora are 37.5 mm. long and 7.5 mm. wide; the tibiae, 37.5 mm. long and 3.7 mm. wide. The tarsi are 15 mm. long, but, since no tarsal claws are visible, it is impossible to determine whether or not the entire tarsus is preserved. The segments of the tarsus are indistinguishable. The femora and tibiae are apparently triangular in section and bear large spines along the inner edges. In this respect the structure of the legs of *Meganeura* is apparently like that of *Typus* (Carpenter, 1939, Fig. 5, p. 37).

SYNONYMY: Brongniart's original type series of *monyi* consisted of five wing fragments and one larger specimen (Fig. 1), which included parts of four wings. However, as stated above, three of these fossils (Figs. 4, 2, 5) are really counterparts of the other three (Figs. 1, 3, 6, respectively) so that the type series actually consisted of but three specimens. Handlirsch, unaware of this, erected five new species on the type series, restricting *monyi* to the specimen illustrated in Figure 1. The species which Handlirsch erected on the type series are as follows:

1. *M. brongniarti* Handlirsch was originally based on the specimens shown in Brongniart's Figures 2 and 3, though Handlirsch did not see the specimens. Lameere pointed out in 1917 that all but two of these characteristics existed in the drawings and not in the fossils; the two differences between *monyi* and *brongniarti* which Lameere regarded as valid were: (a) *monyi* was larger, having a wing length of about 34 cm., which is 4 cm. greater than that of *brongniarti;* (b) in *monyi* the origin of Rs was about 10 cm. from the base, whereas in *brongniarti* it was only about 8 cm. from the base. Both these supposed differences, however, are really due to the error in Brongniart's Figures 2 and 3 on which Handlirsch based his conclusions. As already stated, in these figures the proximal piece of the wing is drawn too near the middle piece, thus shortening the wing as a whole. Furthermore, the basal piece is actually somewhat longer than represented in the figure; in reality Rs arises as the same distance from the base as in the wing shown in Figure 1 (Vignon, 1929, Pl. 2, fig. 4). Measurements of the fossil (not of Brongniart's figures) indicate that the wing of this specimen was about the same size as that represented by Figure 1 (*monyi*). Since there are no real differences between these wings, *brongniarti* is a synonym of *monyi*.

2. *M. brongniartiana* was later erected by Handlirsch (1919) for Brongniart's specimen 2, which he had earlier designated one of the types of *brongniarti*. Actually, specimens 2 and 3 are counterparts of the same fossil, so that *brongniartiana* is a synonym of *brongniarti* and *monyi*.

3. *M. aeroplana* Handlirsch was based on specimen 4, the counterpart of 1, which Handlirsch himself considered the unique type of *monyi*.

4. *M. fafnir* Handlirsch was based on Brongniart's Figures 5 and 6 and characterized by minor venational features, especially in the anal area. As Lameere has

stated (1917), the supposed peculiarities indicated by Handlirsch occur only in the drawings and not in the specimens themselves. A careful comparison of the wing with the hind wings of specimen 1 reveals no difference whatsoever aside from such individual traits as the position of cross-veins. *Fafnir* is undoubtedly a synonym of *monyi*, as it was regarded by Lameere.

5. *M. draco* Handlirsch was based on specimen 5, one of the types of *fafnir*. Since specimens 5 and 6 are counterparts of the same fossil, *draco* is a synonym of *fafnir* and *monyi*.[14]

Meganeurella rapax Handlirsch (1919) was based on a specimen figured by Meunier (1909, Pl. 3, fig. 2), who identified it as *monyi*. I have discussed the structure of this fossil under the synonymy of the genus *Meganeura* above and have pointed out that the specimen shows nothing which distinguishes it from *monyi*. Lameere believed (1917) that this fossil was somewhat smaller than *monyi* and that it belonged to *brongniarti*. However, the dimensions of the preserved part of the wing of Meunier's specimens are very nearly identical with those of the corresponding area in *monyi*. I therefore consider *rapax* a synonym of *monyi*.

Subfamily Typinae

Genus *Meganeurula* Handlirsch

Meganeurula Handlirsch, 1906, Foss. Ins., p. 310; Lameere, 1917, Mus. Nat. Hist. Natur., Bull., vol. 23, p. 161; Handlirsch, 1919, Denkschr. Akad. Wiss., vol. 96, p. 39.
Meganeurites Handlirsch, 1919, Denkschr. Akad. Wiss., vol. 96, p. 60.
Meganeurina Handlirsch, 1919, Denkschr. Akad. Wiss., vol. 96, p. 59.
Gilsonia Meunier, 1909, Mus. Nat. Hist. Natur., Bull., vol. 15, p. 37; Meunier, 1909, Ann. Paleont., vol. 4, p. 143; Lameere, 1917, Mus. Nat. Hist. Natur., Bull., vol. 23, p. 163; Handlirsch, 1919, Denkschr. Akad. Wiss., vol. 96, p. 60.

Species allied to *Typus* and *Megatypus*. Precostal area short, much as in *Typus;* R4 + 5 arising at a point far distal to the origin of Rs; head and thorax like those of *Meganeura,* insofar as known.

Genotype: *Meganeura selysii* Brongniart.

This genus was erected by Handlirsch for *selysii* as a result of his mistaken concept of the structure of R1 in *monyi*. Lameere, having observed this error, related *Meganeurula* to synonymy, but I am convinced that *selysii* differs from *monyi* in at least two characteristics sufficient to require beneric separation—the precostal area and the origin of Rs. In Brongniart's figure the precostal area is as long as that of *monyi;* its small size, however, is readily apparent in the type specimen (Pl. 1, fig. 1), though it is not visible in the photograph by Brongniart. In this respect *selysii* resembles *Typus* more than it does *Meganeura*. It differs from *monyi* also in the late origin of Rs. The venation of the type of *selysii* is not at all satisfactorily preserved, but in the hind wing at least R4 + 5 arises from R at a point remote from the origin of Rs, not close to it as in *monyi*. Brongniart's drawing shows a similar place of origin of R4 + 5 in the fore wing, and, although I cannot discern the actual place of origin of R4 + 5 in fore wing, it undoubtedly arose there as in the hind wing. This late origin of R4 + 5 is most interesting because, together with the precostal area, it has been regarded as the distinguishing feature of the Permian subfamily Typinae. For these reasons, therefore, I believe that *selysii* should be separated

[14] In his catalogue of Paleozoic insects (1921) Handlirsch refers to *draco* as based on Figure 6 and *fafnir* on Figure 5, but this is incorrect.

from *monyi* at least generically. Possibly *selysii* is congeneric with one of the known Permian genera of the subfamily Typinae; but, until we know more about its structure, its assignment to a separate genus seems advisable.

Meganeurites Handlirsch was based on a specimen first recorded and figured by Meunier (1909, Pl. 4, fig 2). The fossil was identified by Meunier as *selysii*, but Handlirsch made it the type of a new species, *gracilipes*. His basis for the genus was given as follows: "In *Meganeurula selysii* Brongniart the legs are very powerful, the hind wing especially strongly broadened. Here (*i.e.*, in *Meganeurites*) just the opposite is found!" Handlirsch's knowledge of this specimen, however, was obtained solely from Meunier's published photograph, which does suggest that the legs were relatively slender. An examination of the fossil shows that this is not really so; the false impression given by the photograph results from the breaking away of the carbonized remains of the legs along the margins, so that the true margins of the legs do not show in the photograph. The dimensions of the legs of Meunier's specimen are in reality identical with those of the type of *selysii*. Since there appear to be no differences in wing structure, either, between *selysii* and *gracilipes*, I am convinced these two species and their genera are synonymous.

Meganeurina was erected by Handlirsch for a specimen mentioned and figured by Meunier as *Meganeura brongniarti* Handlirsch (1909, Pl. 3, fig. 1) but designated the type of a new species (*confusa*) by Handlirsch in 1919. The genus was erected on the assumption that the hind wing was abruptly narrowed basally, as suggested by the published photograph (Meunier, 1909), but the fossil shows that the wing is really only broken away along this line. Unfortunately, very little of the venation is preserved. The late origin of R4 + 5 (left hind wing, as preserved) and the short precostal area (right hind wing), however, show that the insect's affinities are with *Meganeurula* rather than with *Meganeura;* and, although the species is apparently distinct from *selysii* nothing in the fragmentary specimen warrants generic separation.

The genus *Gilsonia* Meunier was erected for *titana* Meunier, based on a nearly complete fore wing. When he formed the genus, *Meunier* had a very erroneous idea of *M. selysii*, and an analysis of his description of *Gilsonia* shows that his concept of the insect agrees almost perfectly with the structure of *selysii* except in size.[15] Unfortunately, the venation of *titana* has been consistently incorrectly described and depicted in the literature. Handlirsch, for example, deriving his knowledge from Meunier's published photograph, mistook the subcosta for the costa and characterized the genus as having a very long and broad precostal area. Actually, the precostal area is very short, R4 + 5 arises at midwing, and all other venational features are similar to those of *selysii*. *Gilsonia* is therefore treated here as a synonym of *Meganeurula*.

Meganeurula selysii (Brongniart)

(Plate 1)

Meganeura selysii Brongniart, 1894, Recherches Hist. Ins. Foss., p. 161, pl. 28; pl. 29, fig. 1; Lameere, 1917, Mus. Nat. Hist. Natur., Bull., vol. 23, p. 161.
Meganeurula selysii Handlirsch, 1906, Foss. Ins., p. 310, pl. 31, figs. 27, 38; Meunier, 1909, Ann. Paleont., vol. 4. pl. 4, fig. 2; Tillyard, 1917, Biol. Dragon-flies, p. 305.
Meganeurites gracilipes Handlirsch, 1919, Denkschr. Acad. Wiss., vol. 96, p. 60; fig. 69.

[15] Since Meunier did not characterize the genus and species separately, his definition of the genus alone cannot be ascertained.

This species is very incompletely known.[16] The fore wing is partially preserved in the type and in one of Meunier's specimens; the proximal third of the wing is almost unknown, and the apical fifth entirely so. The hind wing is represented by the middle third in the type. The largest single piece of either wing preserved is that of the fore wing in the type; it is 137 mm. long and indicates that the complete wing was about 150 mm. long or approximately half the size of that of *monyi*. The fore wing is about 28 mm. wide at midwing, and the hind wing is approximately 35 mm. wide at the same level. The fore wing is fully as slender as that of *Typus* or *Megatypus*. Brongniart's drawing of the wing is very crude, due mainly to the poor preservation of the venation. He has drawn somewhat more of the apex of the fore wing than is actually present in the fossil, and he has also drawn the proximal third of the wing, although, as shown in his photograph of the specimen (Pl. 29), this part of the wing was actually missing. His chief error, however, is in his representation of the precostal area, which he has drawn as long as that in *monyi;* the fossil distinctly shows this to be very small. In the fore wing he depicted Rs arising like an intercalated vein near the origin of M (Brongiart's V); but in the hind wing his Rs (Brongniart's IV) arises at midwing. Handlirsch, in his modified copy of Brongniart's drawing (1906, Pl. 31, fig. 37), corrected the latter error but retained the long precostal area and added a peculiar, falcate tip to the fore wing which is in no way indicated in the fossil. Rs of course arises in both wings as it does in all meganeurids—*i.e.*, at the point of separation of M from R. The most proximal branch of Rs (R4 + 5) can clearly be seen in the hind wing arising at midwing, well beyond the origin of Rs. This is approximately correct in Handlirsch's figure already mentioned. As pointed out under the generic description, the short precostal area and the late origin of R4 + 5 are characteristic of *Typus* and its Permian relatives.

Although Meunier's specimen of *selysii* includes only a small part of the wings, it does show clearly the short precostal area. This fossil is represented by both obverse and reverse. Since the half illustrated in Meunier's paper does not include the procostal area, a photograph of the counterpart, which Doctor Arambourg kindly permitted me to borrow for further study in Cambridge, is reproduced in Figure 2 of Plate 1.

The body structure of *selysii* is better known than that of *monyi*. The type shows the head, thorax, and the two pro- and mesothoracic legs (Pl. 1, fig. 1). The head (without mandibles) is 10 mm. long and 11 mm. wide; the "mandibles" are 7 mm. long.[17] The eyes are almost circular and about 5 mm. in diameter, but they have obviously been flattened considerably. The prothorax is 12 mm. long and 12 mm. wide, the mesothorax 20 mm. long and 16 mm. wide. The metathorax and the abdomen are not preserved. The fore femora and tibiae are about 19 mm. long; the tarsus is 10 mm. long. Brongniart's figure (Pl. 29, fig. 10) shows the tarsus with four segments, but no segmentation was evident when I examined the specimen. The hind legs are almost exactly the same length as the others, the tibiae being 20 mm. long and the tarsus 11 mm.

[16] Meunier has referred to a fine meganeurid which may be *selysii*, as it was identified by him (1914, p. 93). His photograph indicates that this is the best of all the Commentry meganeurids. Unfortunately, I could not find this specimen in the Muséum National, where it was supposed to have been deposited. Meunier does not even give the dimensions of the specimen, and his photograph does not show the venational details needed for generic or specific classification.

[17] The paired appendages which resemble mandibles in this specimen and other Commentry meganeurids probably consist of the maxillae and labium, as well as the mandibles themselves.

Meunier's specimen also includes the head and thorax, as well as part of the abdomen, and it shows many details not preserved in others. The insect has obviously been badly flattened and distorted, however, and apparently both dorsal and ventral structures have been impressed on each counterpart. A photograph of part of this specimen is reproduced in Figure 2 of Plate 1. The head is 10 mm. long and 11 mm. wide (as in the type). The eyes are 6 mm. in diameter and are contiguous, though this may have been caused by the flattening of the head during preservation. The mandibles are broad, toothed, and 4.5 mm. long. A considerable part of the head is posterior to the eyes, but this also may be the result of distortion during preservation. About 10 mm. back of the head is a prominent ridge (r) which resembles superficially at least the ante-alar ridge of the Odonata. Posterior to this is a series of median plates which may represent the tergites of the meso- and metathorax. About 16 mm. back of the ridge is a faint transverse line (t) which I have interpreted as the limits of the thorax. Posterior to this is a suggestion of a median opening (o), which is between two sclerites and vaguely indicative of the accessory genital structures of the second abdominal segments of male Odonata. In the fossil, however, the structures mentioned seem to be on the first abdominal segment. Several abdominal segments are clearly indicated, and, if my interpretation is correct, the first is 4 mm. long, the second 16 mm., and the third 12 mm.; all are about 177 mm. wide. If the other abdominal segments were of similar size the abdomen as a whole would have been a little shorter than the fore wings. The legs are not very well preserved in this specimen; the fore and middle femora and tibiae are identical in length with those of the type of *selysii*. At first glance the legs of this specimen seem more slender than those of the type of *selysii* which induced Handlirsch to erect a new species and genus for the fossil. As already noted, however, more careful examination of the specimen shows that much of the carbonaceous matter of the legs is broken away along the border of the legs; actual measurements show that the legs of Meunier's specimen are as broad as those of Brongniart's type.

Meganeurula confusa (Handlirsch)

Meganeura brongniarti Meunier (nec Handlirsch), 1901, Ann. Paleont., vol. 4, pl. 4, fig. 1.
Meganeurina confusa Handlirsch, 1919, Denkschr. Akad. Wiss., vol. 96, p. 60, fig. 68.

This species is known only by the type, which consists of part of the head, thorax, and the proximal pieces of all four wings. The largest piece of wing is the left hind wing, which is about 125 mm. long. The dimensions indicate that the whole wing was about 200 mm. long or about 50 mm. greater than that of *selysii*. Only a few obscure veins represent the fore wing, but one of the hind wings (left, as preserved) shows the origin of R4+5 at about the same relative position as in *selysii;* the other hind wing shows the precostal area to be very short, even more so than that of *selysii*. The latter wing, incidentally, has two oblique cross-veins between CuP and 1A where the single oblique vein usually occurs in the meganeurids. The posterior margin of the hind wing is broken away proximally, giving the erroneous impression at first glance that the wing is abruptly narrowed basally; this condition misled Handlirsch, who depended upon Meunier's photograph for his information, and induced him to erect a new genus for the specimen.

Very little of the body structure is clear. The head (including mandibles) is 22 mm. in length, which is about 5 mm. longer than that of *selysii*. The legs are more robust than those of *selysii*, being more like those of *monyi*. The fore femora

FIGURE 2. MEUNIER'S SPECIMEN × 1.8
(p. c.) precostal area. For explanation of other lettering, *see* text.

FIGURE 1. TYPE × 1.2

MEGANEURULA SELYSII (BRONGNIART)

and tibiae are 30 mm. long; the middle femora are 36 mm. long, and the tibiae 28 mm. The hind tibiae and tarsi are 29 mm. and 15 mm. long respectively. A comparison of these dimensions with those of *monyi* and *selysii* indicates that the legs of *confusa* were relatively longer than those of either of the other two species; the wings of *confusa* were apparently two-thirds as long as those of *monyi*, but its legs were three-fourths the length of those of the latter species.

Meganeurula titana (Meunier)

(Figure 3)

Gilsonia titana MEUNIER, 1909, Mus. Nat. Hist. Natur., Bull., vol. 15, p. 37, fig. 1; 1909, Ann. Paleont., vol. 4, p. 143, fig. 15; pl. 37, fig. 1; LAMEERE, 1917, Mus. Nat. Hist. Natur., Bull., vol. 23, p. 163; HANDLIRSCH, 1919, Denkschr. Akad. Wiss., vol. 96, p. 61, fig. 70; VIGNON, 1929, Mus. Hist. Natur., Arch., vol. 4, pl. 2, fig. 3.

This species is based upon a well-preserved fore wing, lacking only about the distal third of the posterior border. As preserved, the wing is 11 mm. long, and the nature of the venation distally indicates that the complete wing was about 120 mm. long. This is 30 mm. smaller than the wing of *selysii*. The precostal area is very short, and there is a slight bend in the contour of the anterior wing margin at the distal end of this area; the subcosta extends to the end of the preserved part of the wing, which is obviously not far from the true apex. Rs arises from M a very short distance from the point of origin of the latter, but this is almost certainly an individual trait, since a similar origin of Rs occurs in some specimens of other Meganeuridae (*Typus*); MA, CuP, and 1A are also formed as in *Typus;* the area above CuP basally is broken away so that it is impossible to determine whether or not the supposed vestige of CuA was present as in *Typus*. The cross-veins are arranged as in *Typus,* forming cellules in the anal area; between 1A and CuP there are two oblique cross-veins proximally; there is no sign of the sclerotized area at the base of the anal area as in *Meganeura,* only a weak, oblique vein like that in *Typus permianus.*

The venation has been variously and inconsistently described in the literature. Meunier's original drawing is very crude and definitely incorrect in several respects, such as the origin of Rs and MA. Lameere, who stated that the fossil was well represented by Meunier's drawing, placed much emphasis on the apparent origin of Rs from MA, just beyond the origin of the latter; such an origin of Rs, however, is almost certainly an individual trait; it occurs in some individual wings of *Typus permianus* and not in others and may be brought about by a shift of only about a millimeter in the point of origin of Rs from R+M. Handlirsch's figure of *titana* (1919, Fig. 70) is absurd; it depicts the precostal area as extending beyond midwing and as being relatively larger and longer than that of *Meganeura*. Of course his drawing and description of the genus were based on Meunier's published photograph, which was not very clear. The genus *Gilsonia* has therefore been erected and maintained on characteristics which unquestionably do not exist in the specimen.

The venation of *titana* is much better known than that of *selysii,* and similarities between its venation and that of *Typus* are striking. When the wings of *Meganeurula* are better known, especially those of the genotype, separation of this genus from *Typus* may be untenable.

Composition of the family Meganeuridae.—The family Meganeuridae is usually divided into two subfamilies, Meganeurinae and Typinae, though Handlirsch has suggested the elevation of Typinae to family rank. In the Meganeurinae the

precostal space is very long, extending to about the middle of the wing, and R4 + 5 rises close to the origin of Rs; in the Typinae the precostal space is short, being confinded almost to the base of the wing, and R4 + 5 arises about at midwing, far from the origin of Rs. All the Commentry meganeurids have previously been classified in the Meganeurinae, and, except for *Meganeuropsis* (Carpenter, 1939), all the Permian meganeurids have been considered Typinae. As already pointed

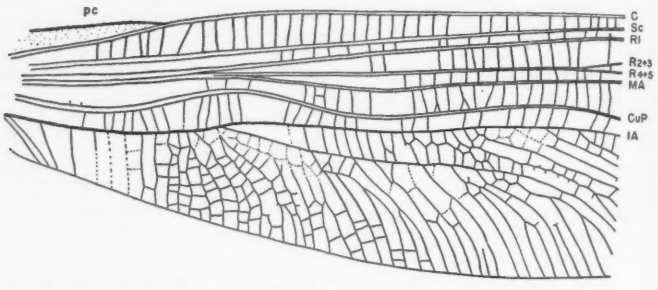

FIGURE 3.—*Proximal part of wing of* Meganeurula titana *Meunier*
Original figure, drawn from type. (R2+3, R4+5) branches of radial sector (—); other lettering as in Figure 1.

out, however, *Meganeura monyi* is the only Commentry species really belonging to the Meganeurinae; all the others belong to the Typinae. Both these subfamilies therefore existed in the Carboniferous and Permian.

In my account of the Protodonata from the Permian of Kansas (1939, p. 43), I included a table summarizing my concept of the order. The observations which I have subsequently made on the Commentry types require several changes in that concept, notably the elimination of the Protagriidae and Calvertiellidae and the recognition of the Typinae in the Upper Carboniferous. A revised synopsis of the order follows:

Order Protodonata

1. Family Meganeuridae (= Meganisoptera Martynov)
   A. Subfamily Meganeurinae Handlirsch. Upper Carboniferous of Europe. Permian of Kansas, Oklahoma.
   B. Subfamily Typinae Handlirsch. Upper Carboniferous of Europe. Permian of Kansas, Arizona, Oklahoma, Russia.
2. Family Paralogidae Handlirsch. Upper Carboniferous of North America (Rhode Island).
3. Family Erasipteridae Carpenter. Upper Carboniferous of Europe (Czechoslovakia).

FAMILY CAMPYLOPTERIDAE (ORDER MEGASECOPTERA)

Systematic descriptions.—This monotypic family is based on *Campyloptera eatoni* Brongniart from Commentry. Since only one wing is known, its affinities are uncertain, but it has generally been referred to either the Megasecoptera or the Protodonata. Careful examination has shown that more of the venation is really preserved than has been supposed. The additional characteristics now known, however, seem to confuse rather than to clarify its systematic position. A distinct

order will probably be needed eventually for the family, but, until more is known of the insects, I believe it should be referred to the order Megasecoptera. The new suborder Campylopteroidea is here established for the family to emphasize its divergence from the true Megasecoptera (Eumegasecoptera) and the related Protohymenoptera, neither of which possesses intercalated sectors. It is impossible to indicate characteristics of the family Campylopteridae until other Campylopteroidea are known; probably the nodal bend of the costal margin, the petiolation of the wing, and the pterostigma will fall in this category.

Genus *Campyloptera* Brongniart

Campyloptera BRONGNIART, 1894, Recherches Hist. Ins. Foss., p. 406.

Wings slender; Sc terminating at about midwing; a nodal bend similar to that of the Odonata is present; R1 terminating at apex; pterostigma apparently present; MA anastomosed with Rs proximally.

GENOTYPE: *Campyloptera eatoni* Brongniart

Campyloptera eatoni Brongniart

(Figure 4)

Campyloptera eatoni BRONGNIART, 1894, Recherches Hist. Ins. Foss., p. 406; pl. 24, fig. 3; HANDLIRSCH, 1906, Foss. Ins., p. 316, pl. 32, fig. 15; TILLYARD, 1928, Ind. Mus., Rec., vol. 30, p. 161; LAMEERE, 1917, Mus. Nat. Hist. Natur., Bull., vol. 23, p. 163.

This species is based on a single, poorly preserved wing, represented by obverse and reverse. The latter half of the fossil has been coated with a substance resembling shellac, now so dark as to obscure nearly all the veins. The obverse, fortunately, has not been so treated, and under good illumination most of the veins can be distinguished. The specimen was studied with the aid of photographs of the fossil enlarged to about 2 feet in length; in this way many venational details not previously described were observed.

The wing, apparently complete, is 64 mm. long and 11 mm. wide. Its margins can be followed without difficulty except for the proximal part of the posterior border. A careful examination of the fossil under high binocular magnification reveals the margin even here; it is close to the stem of 1A and thus forms a short but distinct petiole to the wing. The very base of the anterior margin is curved inward. As Lameere noted (1917), there is no indication of a precostal area, though Brongniart showed one in his figure of the specimen. The anterior margin of the wing is not straight or smoothly curved, as drawn by Brongniart; it is distinctly bent at the termination of Sc, somewhat proximal to the midwing. The apex is slightly pointed. A small piece of the wing margin is broken away directly above the apex; proximal to the broken section there is a small dark area that appears to be part of a pterostigma. The subcosta is straight and is connected to the anterior margin by about 8 cross-veins, the proximal one of which is much stronger than the others and is in line with the one below it, thus resembling the primary antenodal of the Odonata.[15] R and M are probably anastomosed proximally; the origin of MP from R+M is not preserved, but MA is anastomosed with Rs for a considerable distance before it separates as an independent vein; R1 terminates at the apex of the wing; Rs gives rise to two branches; R4 + 5 arises not far from the point of separation of M from Rs; R2 and R3 diverge just beyond midwing; between R2

[15] A similar alignment corresponding to that of the second antenodal may have been present a little farther distad, but that particular part of the wing is broken away.

and R3 is a convex intercalated vein (IR2), and a similar one (IR3) occurs between R3 and R4 + 5; MA is unbranched and terminates in the distal third of the wing; the base of MP, which is strongly concave, is not preserved; it may arise from either R+M or Rs+M; the base of Cu is also missing; CuA and CuP are discernible from about the level of the origin of Rs+MA; both are unbranched; 1A is preserved to the base of the wing; that portion of it in the periole is slightly curved; it is unbranched but connected to the posterior margin by several cross-

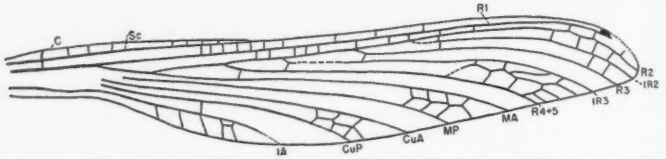

FIGURE 4.—*Wings of* Campyloptera eatoni *Brongniart*

Original figure, drawn from type. (IR2, IR3) intercalated veins (+) of radial sector; other lettering as in Figure 1.

veins. Cross-veins show between the main veins of the wing; between some veins distally (R4 + 5 and IR3, MA and MP, CuA and CuP) are two rows of cells.

It is most unfortunate that this remarkable fossil is not better preserved proximally; it is one of the most interesting species in the Commentry fauna. Opinions have differed regarding the venation of *Campyloptera,* but I believe the structure of most of the debated features is definitely shown in the fossil. Brongniart's figure of the specimen is accurate in general, but not so in many details, a few of which are of phylogenetic importance. He indicates with a faint line a prominent precostal area; this is reproduced in Handlirsch's figure (copied from Brongniart) and is mentioned by Tillyard, who did not see the fossil, as one of the significant characteristics of the insect (1928, p. 160). Lameere, however, after examining the specimen, stated (1917, p. 163) that there was no precostal space. I agree with this statement; an examination of the specimen under various illuminations and magnifications failed to reveal any sign of it.

Brongniart's figure shows the apex of the wing much more pointed than it actually is. He apparently failed to observe that a small piece of the anterior wing margin was broken away just above the apex, thus giving the erroneous impression that the apex was very pointed. He did not, of course, mention the convexity or concavity of the veins, though these are well indicated in the fossil and are readily observed in photographs of the wing taken under oblique illumination. On the basis of Brongniart's figure, Handlirsch was led to believe that there were intercalated veins in the region of Rs, and his opinion was emphasized by Tillyard in his discussion of the affinities of *Campyloptera.* Lameere (1917), however, following his examination of the specimen, asserted that intercalated sectors were not present. I have previously accepted Lameere's statement as correct (1931) but, from subsequent examination of the fossil, I am convinced that intercalated veins do exist. They are so well indicated by their convexity that they show clearly in photographs.

Affinities of Campylopteridae.—The relationships of *Campyloptera* have been the subject of much controversy. Brongniart originally (1885)[10] placed the genus

[10] Although the generic name was used by Brongniart in 1885, no species was mentioned then, and *Campyloptera* remained a *nomen nudum* until 1894.

in the Megasecoptera but later (1894) assigned it to the Protodonata. Handlirsch referred it back to the Megasecoptera in 1906, but Tillyard in 1928 again transferred it to the Protodonata and even the family Protagriidae. Lameere (1917), after a study of the fossil, concluded it was a megasecopteron, but, because of the poor preservation of the specimen, he made no attempt to determine its affinities within the order.

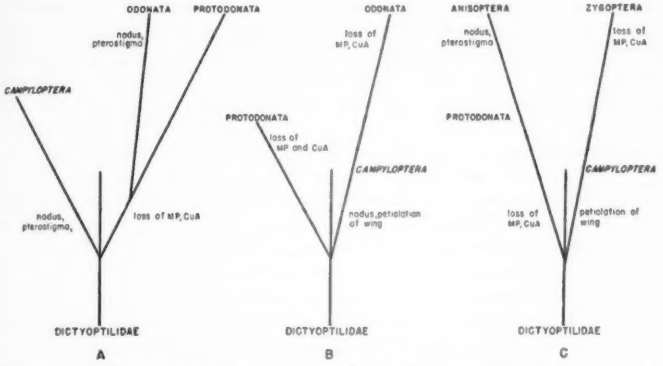

FIGURE 5.—*Diagrams indicating possible positions of* Campyloptera *in phylogeny of Odonata and Protodonata*

For explanation, *see* text.

Two chief reasons for the difference of opinion about the affinities of *Campyloptera* are: (1) the poor preservation of the unique specimen and the consequent unsatisfactory figure given by Brongniart; (2) the venational peculiarities of the insect. The former is now not nearly so important as the latter, for, as already mentioned, most of the wing veins really are preserved and can be determined with certainty. Figure 4 is a drawing of the wing including only those structures which I am convinced are present and which can be seen in photographs of the specimen. Many features shown in this drawing were not even suggested in Brongniart's and are of phylogenetic interest: (1) absence of a precostal area; (2) presence of a short Sc and nodal bend in the anterior margin of the wing; (3) nature of the cross-veins, especially the presence of a very strong cross-vein in the subcostal area; (4) probable (but not certain) presence of a pterostigma; (5) coalescence of MA with RS and the probable coalescence of the stem of M with R; (6) presence of two convex intercalated veins in the region of Rs; (7) presence of MP (—) and CuA (+); (8) presence of a single anal vein; (9) petiolation of the wing.

In attempting to determine the affinities of *Campyloptera*, we need concern ourselves only with the palaeopterous orders; of these we can surely eliminate at once the Protephemerida and Plectoptera, both of which have a very generalized venation. The orders to which *Campyloptera* might conceivably belong are: (a) Megasecoptera, (b) Odonata, (c) Protodonata, and (d) Palaeodictyoptera. (a) The Megasecoptera and *Campyloptera* have several features in common, such as the petiolate wing, pterostigma, complete MP and CuA, and the coalescence of Rs and MA. None of the Megasecoptera, however, have a nodal bend or intercalated sectors. (b) The odonate features of *Campyloptera* are very striking, especially those of the zygop-

terous Odonata; the similarities are not only in such general traits as the petiolation of the wing but also in details, as the nodal bend, the presence of an apparent primary antenodal, and of the intercalated veins IR2 and IR3. In fact, all but one of the phylogenetic features listed are characteristic of at least the zygopterous Odonata. The exception is the presence of MP and CuA in *Campyloptera* and their absence in all Odonata. This is a most important difference and certainly requires the elimination of *Campyloptera* from the order Odonata itself; (c) the possible relationship between *Campyloptera* and the Protodonata depends on how the latter is defined; none of the insects generally placed in the Protodonata have a nodus or petiolate wing; and as the order has been defined here (equivalent to the Meganisoptera) *Campyloptera* would be eliminated by the presence of MP and CuA. (d) In comparison with the typical members of the Palaeodictyoptera, *Campyloptera* is extremely specialized; so far as known none of the insects thus far placed in this group have a nodus, pterostigma, intercalated sectors, or even petiolate wings. *Campyloptera* will therefore not fit readily into any of the orders thus far established.

In view of the striking similarities between *Campyloptera* and zygopterous Odonata, and especially because of Fraser's conclusion (1938) that the latter were ancestral to the Anisoptera, I have considered the positions *Campyloptera* might have held in the phylogeny of the odonate complex. The probable relationships among the Protodonata—*i.e.*, Meganisoptera, Palaeodictyoptera, and Odonata—have been generally agreed on. A palaeodictyopteron of a dictyoptilid type (having MA, MP, CuA, CuP) apparently gave rise to the Protodonata and the Odonata. Although the Odonata were probably not derived from any family of Protodonata so far known, these two orders were almost certainly from a common ancestral stock. Compared with the Palaeodictyoptera, both are specialized in the absence of MP and CuA; and the Odonata are more specialized than the Protodonata in having at least an incipient nodus. The three possible positions which *Campyloptera* could have occupied in this odonate phylogeny are shown in Figure 5. If *Campyloptera* held any one of the positions indicated, there has been an extraordinary convergence in the evolution of the odonate complex.

Inasmuch as we do not know anything about the body structure of *Campyloptera* or even whether the type is a fore or hind wing, its systematic position must temporarily remain obscure. In the meantime, rather than attempt to fit it in the odonate complex or the Palaeodictyoptera, it seems advisable to place it in a separate suborder (Campylopteroidea) of the order Megasecoptera, from the other members of which it differs less than it does from the members comprising the odonate groups.

SUMMARY

Study of the Commentry insects in the Muséum National d'histoire Naturelle shows that:

(1) The family Protagriidae is more closely related to the Palaeodictyoptera than to the Protodonata and is here assigned to the former order. It is only very remotely, if at all, related to the Odonata.

(2) The family Meganeuridae includes the typical Protodonata. Only four valid species occur in the Commentry shales, the seven other described species being synonyms.

(3) The family Campylopteridae is an aberrant and highly specialized group, whose characteristics are suggestive of the Protodonata, Odonata, and Megasecoptera. It is tentatively placed in a new suborder (Campylopteroidea) of the Megasecoptera until additional characteristics are known.

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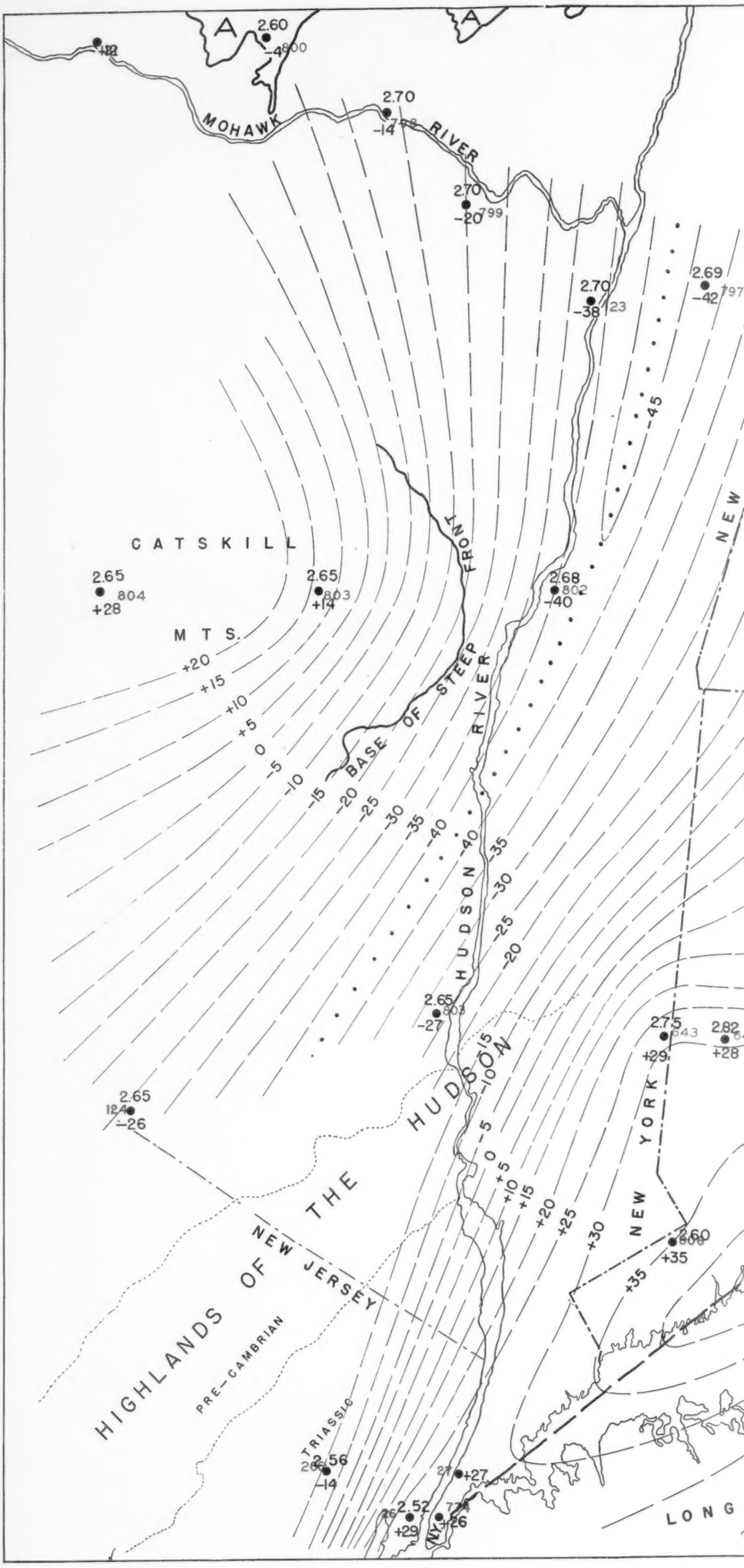
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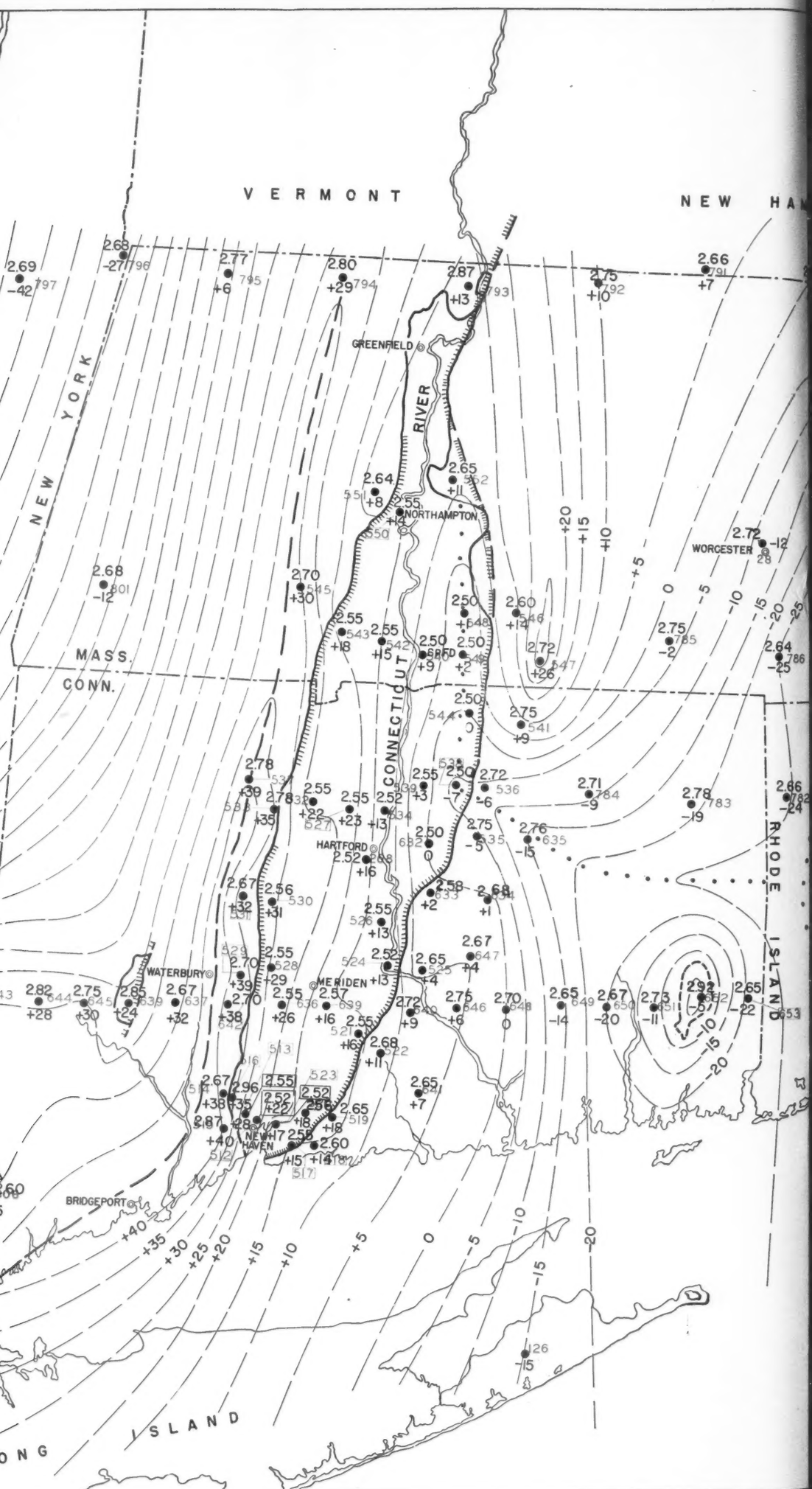
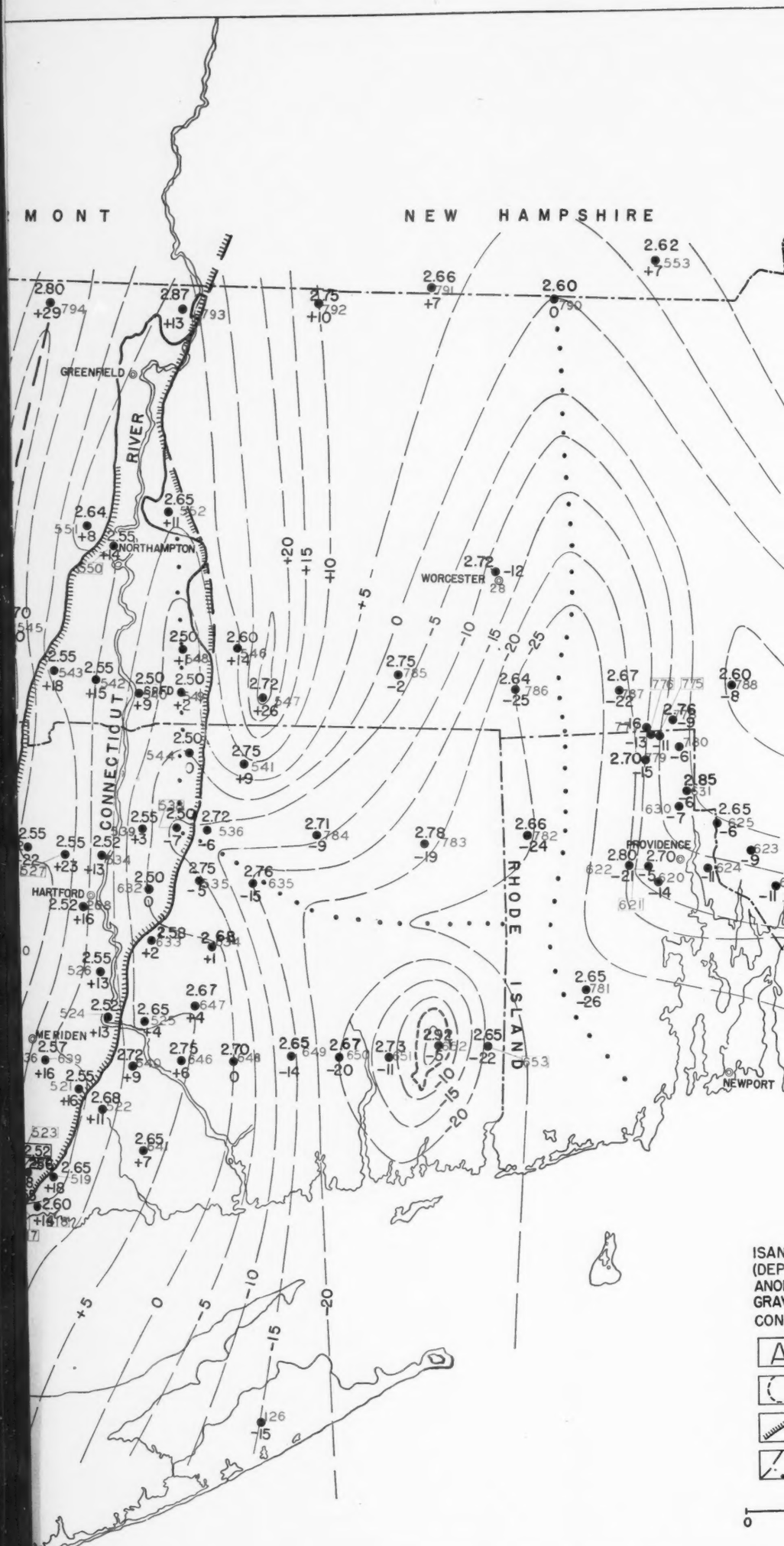


ISANOMALY MAP, BASED ON ISOSTATIC ANOMALIES



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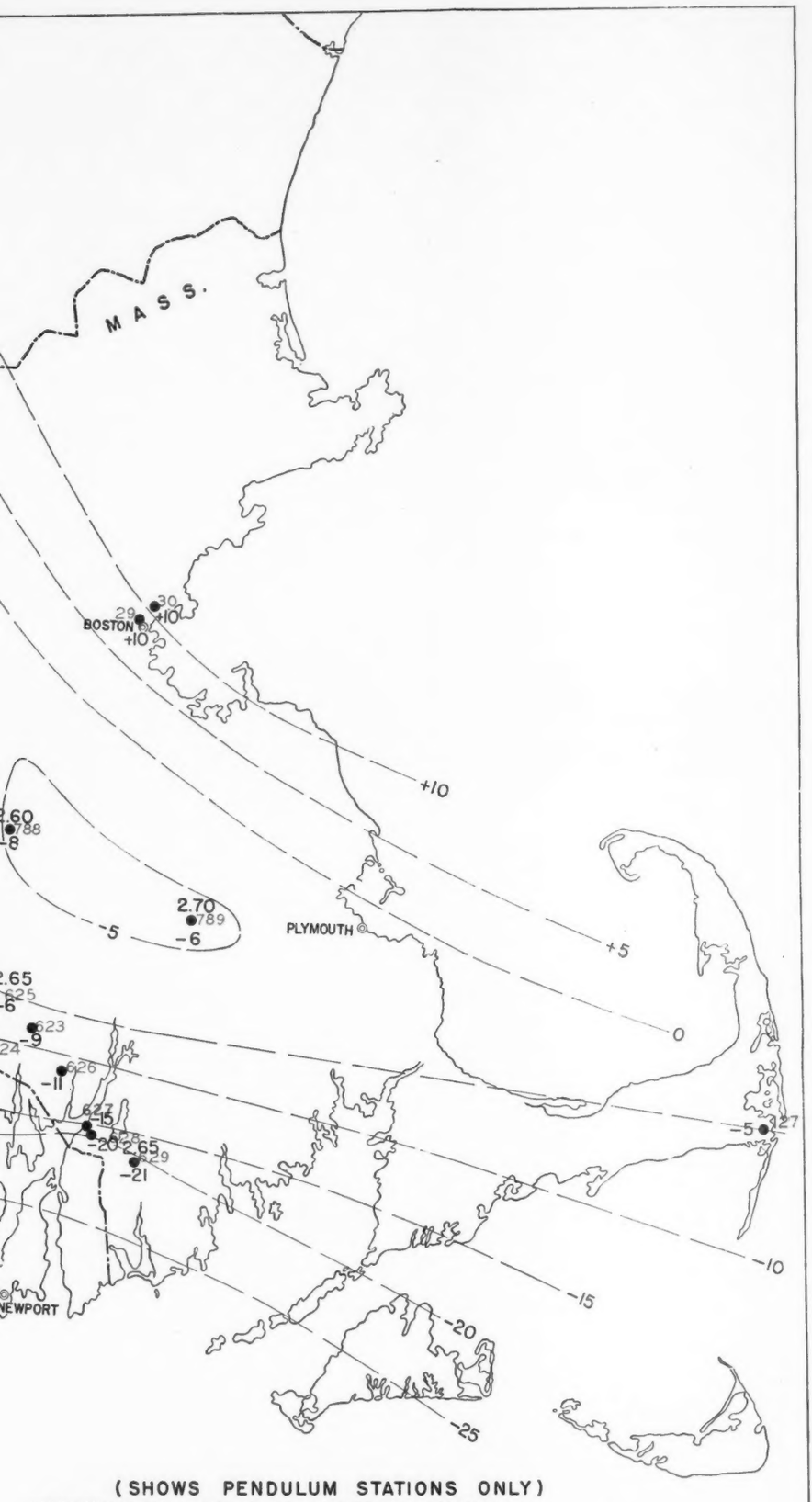

(SHOWS PENDULUM STATIONS ONLY)

ISANOMALY MAP, BASED ON ISOSTATIC ANOMALIES (DEPTH OF COMPENSATION, 56.9 KM.). FIGURES SHOW ANOMALIES IN MILLIGALS, AND AVERAGE SPECIFIC GRAVITIES OF SURFICIAL ROCKS AT STATIONS. (STATION NUMBERS IN RED) CONN-MASS. TRIASSIC AREAS INCLOSED WITH HEAVY LINES.

A — PARTS OF ADIRONDACK MOUNTAINS.

OUTCROP OF PRESTON GABBRO.

FAULT (HACHURES ON DOWNTHROW SIDE). BROKEN LINE INDICATES PROBABLE POSITION OF FAULT.

AXES OF GRAVITY "HIGHS" (BROKEN LINES) AND "LOWS" (DOTTED LINES).

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GEOLOGIC INTERPRETATION OF GRAVITY ANOMALIES IN THE SOUTHERN NEW ENGLAND–HUDSON VALLEY REGION

BY CHESTER R. LONGWELL

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ABSTRACT

Pendulum stations occupied by the U. S. Coast and Geodetic Survey at locations chosen for their geologic significance provide a net in southern New England, eastern New York, and northern New Jersey comprising about 120 stations. Numerous supplemental stations have been added with a modern gravimeter. Specific gravities of representative rocks near the pendulum stations have been determined.

Isanomaly maps based on isostatic, Bouguer, and free-air anomalies are remarkably alike in their major features. Some groups of anomalies reflect bedrock units that have exceptional densities. When corrections are applied for these local masses, regional tendencies are clarified. Two belts of negative anomalies are separated by a belt of positive anomalies. The negative belts correspond in a general way to axes of subsidence during Paleozoic periods. The entire region experienced orogenic deformation, followed by peneplanation.

Since the groups of Bouguer anomalies are little affected by the ordinary isostatic corrections, and are related only in minor degree to the visible bedrock, it appears that they either indicate considerable departures from isostatic balance or reflect differences in density that are concealed. Glennie's hypothesis of crustal warping, in a modified form, offers an attractive explanation of the outstanding anomalies.

INTRODUCTION

STATEMENT OF THE GENERAL PROBLEM

Early in the quantitative study of isostasy it was realized that variations in density of rock masses at or near the Earth's surface are an important cause of differences between observed and computed gravity values. A pendulum or a gravimeter is affected not merely by the total mass in the rock column beneath it, but also by the distribution of mass. A large salt plug directly under the instrument makes the measured value of gravity less than the theoretical value even if the defect in mass is compensated at greater depth, because gravitative attraction varies inversely as the square of the distance. Hayford and Bowie suggested near the beginning of their studies that some negative anomalies can be partly explained by the presence of poorly consolidated young sediments, and some positive anomalies by bodies of exceptionally dense rock, beneath the stations concerned (Hayford, 1910; Hayford and Bowie,

1912; Bowie, 1917). David White was the first geologist to exploit this idea in an exhaustive study of gravity values in the United States considered in their relation to the local geology (1924). More recently geologists have given increased attention to gravity anomalies as possible effects of lithology and structure. The concept has had important application in the commercial field, particularly in the search for conditions favorable for the occurrence of petroleum. Chamberlin's analysis of anomalies for the Black Hills-Big Horn-Beartooth region (1935) marks an advance in adaptation of the principle as a scientific aid for attack on regional structural problems.

PRESENT INVESTIGATION

About 10 years ago Adolph Knopf and the writer discussed with Bowie, who was then Chief of the Division of Geodesy, U. S. Coast and Geodetic Survey, the desirability of establishing a network of gravity stations on and near the Connecticut Valley Triassic rocks, in Connecticut and Massachusetts. At that time the only station in this area was No. 268, at Hartford. This large Triassic unit, a comparatively simple wedge with average subnormal density, faulted into a basement of distinctly higher density, appeared to present an exceptional opportunity for a quantitative check of the degree to which gravity values reflect the local geology. In the summer of 1934 a gravity party, under H. E. McConnell, was assigned to start a co-operative program in New England. During that summer 42 stations were established on and near the Triassic area, at locations carefully selected for their geologic significance. In 1935 the network was extended, by a party under B. R. Wohld; 22 additional stations were occupied, chiefly in Connecticut, and 12 stations were established near Providence, Rhode Island, at locations selected by Charles W. Brown. Preliminary study of the anomalies brought out several points of geologic interest but also showed the need of additional stations to extend the regional picture. In 1937 a Coast and Geodetic Survey party under C. A. Schoene added 6 new stations to the network about Providence and occupied 26 stations at locations selected by the writer in Connecticut, Rhode Island, Massachusetts, southern New Hampshire, and eastern New York. The 3-year program yielded a total of 108 new pendulum stations; these, with 12 older stations in the region, form the basis for the isostatic isanomaly map (Pl. 1) and the gravity profiles (Figs. 6, 7, 8, 9) presented in this paper.

While the study was under way, announcements were made of rapid and accurate measurements of gravity by new sensitive gravimeters. Publication on the New England work was delayed in the hope that a gravimeter might be secured to fill some obvious gaps in the network

of stations. George P. Woollard, using an instrument loaned by the Humble Oil and Refining Company, occupied 70 supplemental stations in the region late in the season of 1940 and added about 100 more in the Summer and Fall of 1941. He made Bouguer and free-air anomalies available to the writer while the present paper was in final preparation. Incorporation of Woollard's data would necessitate considerable modification of all the isanomaly maps, especially in the eastern and southern parts of the map area. Since isostatic values have not yet been computed for the gravimeter stations, Plate 1 does not show these stations but presents isanomaly lines inferred from the pendulum stations only. Free-air isanomalies based on pendulum values are retained in Plate 3, but Woollard's stations of 1940 are added to this chart. All the 1940 and 1941 gravimeter values are incorporated in the Bouguer chart (Pl. 2), and isanomaly lines on this chart have been redrawn to take account of all available stations.

Specific gravities of rocks at the gravimeter stations are not available; they have been determined for nearly all the pendulum stations.

ACKNOWLEDGMENTS

The writer is especially indebted to officers of the U. S. Coast and Geodetic Survey, whose full co-operation made the present study possible. George P. Woollard generously contributed new field and office data. Parts of the material have been presented orally before several groups of geologists and geophysicists (Longwell, 1937b; 1938), whose critical discussions were a valuable aid in bringing the study to its present stage.

GENERAL GEOLOGY OF THE REGION

GENERAL STATEMENT

The reader will be able to evaluate an analysis of the gravity anomalies only if he has some knowledge of the major geologic features of the region. A brief outline of the areal geology and of the geologic history is presented here; for greater detail the reader is referred to the references cited and to published geologic maps of New York, Connecticut, Massachusetts, and Rhode Island.

PALEOZOIC SYNCLINORIUM OF HUDSON VALLEY

Sedimentary formations ranging from Cambrian to Devonian occupy a belt that roughly parallels the upper part of the Hudson River and swings southwestward along the west side of a pre-Cambrian block forming the Highlands of the Hudson (Pl. 1). Some of these formations extend into western New England, but the eastern limit of the Paleozoic belt is not known; age relations become obscure in a zone of intense

deformation and metamorphism that extends generally north-south in Connecticut, Massachusetts, and eastern New York. During the several Paleozoic periods indicated above, a geosynclinal trough extended northeastward from the Appalachian region along the east side of the Adirondack mass (Schuchert and Dunbar, 1941). The axis of maximum sedimentary accumulation seems to have been in the Hudson Valley belt, and the prism of deposits was further thickened in this belt by folding and thrusting near the close of the Ordovician and also later in the Paleozoic.

METAMORPHIC AND IGNEOUS ROCKS OF SOUTHERN NEW ENGLAND

The New England States are underlain in large part by metamorphic rocks that are difficult to interpret at best and have not been satisfactorily dated in large areas. Quartzite and metamorphosed carbonate rocks in western Massachusetts and northwestern Connecticut are related to the early Paleozoic sequence in the Hudson Valley belt. Cambrian slates in eastern Massachusetts represent a sedimentary section that probably was many thousands of feet thick. Many of the schists and gneisses in Connecticut and Massachusetts were derived from sedimentary rocks of unknown ages, although some of the units have been assigned doubtfully to the Ordovician and Silurian. Devonian fossils have been found in northern Massachusetts, and in New Hampshire beyond the limits of the area considered here. Thick Pennsylvanian deposits, in part coal-bearing, extend from Rhode Island into eastern Massachusetts; conglomerates and slates assigned to the Permian occur in a limited area near Boston. These late Paleozoic formations, all nonmarine, appear to have been derived from a land mass that lay east of the present shoreline. The sediments were laid down in a subsiding trough that occupied generally the position of an earlier geosyncline (the Acadian trough of Schuchert), in which marine sediments accumulated during the early Paleozoic periods. This trough was parallel to the main Appalachian trough in the Hudson Valley Region. The formations in the eastern trough were deformed in the Appalachian revolution; the older part of the section probably was involved also in earlier Paleozoic orogenic movements (Emerson, 1917).

Numerous intrusive igneous masses, ranging in composition from granite to gabbro, cut the metamorphosed formations. Many of these masses made important contributions to the metamorphic effects. By use of stratigraphic evidence, some of the granitic bodies have been dated as post-Pennsylvanian. On the basis of radioactive minerals, at least one granitic mass in Connecticut is assigned tentatively to the late Ordovician,

and several others are believed to be Devonian. Pre-Cambrian gneisses, in part of igneous origin, are recognized in several areas.

Field workers in New Hampshire and Vermont report thick sections of metamorphosed sedimentary strata east of the Green Mountains (Billings, 1937; Jahns and White, 1941). No gravity data are available for that region, which therefore is excluded from the present study.

TRIASSIC BEDROCK

A large area underlain by Triassic rocks extends northward from New Haven entirely across Connecticut and almost to the northern boundary of Massachusetts (Pl. 1). This area is more than 100 miles long, and its greatest width is 21 miles. The Triassic sedimentary rocks consist of continental arkose, sandstone, siltstone, and shale, predominantly red and brown, exceptionally gray or black. Three extensive basaltic flows lie somewhat above the middle of the sedimentary section; dikes and sills chemically like the flows are distributed irregularly in the Triassic area. Large dikes of similar composition and presumably of Triassic age cut the older rocks both east and west of the Triassic belt.

In general the Triassic beds dip homoclinally eastward and have been beveled by erosion (Fig. 1). Some faulting has occurred along the western border, but in some stretches of that border the basal Triassic is in sedimentary contact with the older rocks. A great normal fault marks much of the eastern border, and the maximum thickness of the Triassic section near this fault is supposed to be nearly 3 miles.

A much smaller area of Triassic rocks lies in the Pomperaug Valley, southwest of Waterbury (Pl. 1). This unit resembles the larger Triassic belt in lithology and structure: red beds including two basaltic sheets dip toward an eastern boundary fault. The maximum thickness of the section is about 1200 feet.

The northern part of the large New York-New Jersey Triassic area lies west of New York City, with its eastern boundary along the Hudson River (Pl. 1). Sedimentary beds similar to those in Connecticut and Massachusetts, including sheets of basaltic lava with associated dikes and sills, dip moderately westward toward a fault contact with pre-Cambrian rocks in the Highlands. The thick Palisades sill of dolerite lies near the base of the section, directly west of the Hudson. Only two of the pendulum stations and eight of the new gravimeter stations are on the part of this Triassic area considered in the present paper (Pl. 2).

COASTAL-PLAIN STRATA

On Long Island outcrops of Cretaceous clays and sands seem to continue beneath the widespread veneer of glacial debris, as indicated by numerous well records in all parts of the island (Fuller, 1914). These

Coastal-Plain deposits are only a few hundred feet thick at a maximum. Probably some remnants of Tertiary deposits lie on the Cretaceous locally; the evidence comes from lithologic comparison with known Tertiary formations in nearby New Jersey. A gentle dip toward the southeast characterizes the Cretaceous beds on Long Island, although the weak beds exposed at the surface have been crumpled and faulted, presumably by pressure from the Pleistocene glacier ice.

No Coastal-Plain deposits have been identified on the mainland north of Long Island, though they extend eastward along the continental shelf and form limited outcrops in the coastal belt of southeastern Massachusetts. Cretaceous fossil wood in glacial drift in Connecticut, nearly 20 miles north of the Sound, suggests strongly that a Cretaceous cover once extended far inland.

PLEISTOCENE DRIFT

Till and stratified drift mantle a large part of the surface to varying depth. Apparently unconsolidated surficial deposits are nowhere thick enough to have an appreciable effect on gravity values. Around some of the stations, however, particularly in the eastern part of the Connecticut Valley Triassic area, the bedrock is so effectively covered with drift that it is difficult to collect representative specimens for specific-gravity determinations.

STRUCTURAL TRENDS OF THE REGION

Faults, fold axes, and schistosity strike generally north or somewhat east of north. There are numerous local exceptions, near the borders of intrusive igneous bodies and in connection with the pitching of major folds. However, the northerly trends dominate the regional "grain," which was impressed on the bedrock by the major diastrophic movements. The larger dikes of pegmatite and of dolerite, formed in widely separated epochs, parallel the dominant structural trend, as do the large normal faults affecting the Triassic beds.

OUTLINE OF GEOLOGIC HISTORY

Many important events in the history of the region are obscure or unknown, because in large areas the sedimentary record is either lacking or interrupted by large gaps. Severe metamorphism, both kinetic and thermal, has affected most of the pre-Triassic rocks, except in some of the Paleozoic formations west of the Hudson. In large areas it is not possible to say whether schistose and gneissic rocks are pre-Cambrian or Paleozoic. Insofar as the history can be deciphered, a few major facts seem to have some bearing on the interpretation of gravity values.

During the early Paleozoic periods there were two subsiding tracts receiving thick sedimentary deposits in this general region (Schuchert and Dunbar, 1941). The St. Lawrence trough, with its axis in the general position of the present Hudson Valley, was separated by a land mass—named by Schuchert the New Brunswick geanticline—from the Acadian trough which extended northeast through Rhode Island and eastern Massachusetts. Formations in the St. Lawrence trough, which was the northern extension of the main Appalachian geosyncline, were strongly deformed during at least two widely separated epochs (Schuchert and Longwell, 1932): the first in late Ordovician time, the other probably in the late Paleozoic Appalachian revolution. The position of the New Brunswick geanticline, extending somewhat east of north across central Connecticut and Massachusetts, seems fairly well defined from early Paleozoic time through the Silurian (Schuchert and Dunbar, 1941), but less definite thereafter. Probably its outline was modified by the Taconian orogeny (late Ordovician), which intensified a positive unit somewhat farther to the west, the Green Mountain axis, extending southward from Vermont. This orogeny deformed the strata in the St. Lawrence trough, and for a time following the disturbance the western margin of the uplifted mass lay well to the west, within the geosyncline, as indicated by progressive eastward overlap of Silurian beds across closely folded and eroded Ordovician formations in that district.

Unfortunately there is little evidence to fix the position of the axis of uplift in post-Silurian time. Further uplift affected an unknown extent of the New England region in the Devonian period; the resulting highland became the source of the thick Catskill clastic sediments that accumulated over a wide region west of the present Hudson Valley. This Devonian uplift probably was related to the Acadian orogeny, which affected a large region in southeastern Canada and New England, although its southward extent has not been determined. The third Paleozoic orogeny of the region—the Appalachian revolution of post-Pennsylvanian date—deformed the late Paleozoic formations in Rhode Island and eastern Massachusetts; its possible effects in a large part of the region cannot at present be differentiated from effects of the earlier orogenies. Regional uplift accompanied or followed the orogeny, but there is no evidence bearing on the particular behavior during that time of the old axis in western New England. During the Triassic period there was uplift and depression along steep faults in the Appalachian region, with local accumulation of thick sedimentary deposits on sinking blocks. Outpouring of basaltic lava accompanied the faulting and sedimentation. The great fault along the eastern margin of the Connecticut Valley Triassic area was active in Triassic time. A

rising block east of the fault contributed sediments to a sinking block west of it; a wedge-shaped deposit, thickest near the fault, extended westward beyond the Pomperaug Valley area (Longwell, 1937a). Later there was strong uparching in western Massachusetts and Connecticut, as indicated by the present eastward homoclinal dip of the New England Triassic strata and the westward dip in the New York-New Jersey area. This arching uplift, accompanied by large-scale faulting, suggests rejuvenation along the old Green Mountain axis.

After the Triassic period the region was reduced to low relief. A thin veneer of Coastal-Plain sediments perhaps encroached far inland but was eroded from the mainland following broad regional upwarping during the Tertiary. Since the uplift many valleys have been cut, lowlands have been developed on the weaker bedrock units, and many parts of the region have acquired moderate relief. However, there is no high mountainous topography with abrupt relief, a favorable circumstance in the kind of study attempted here. Only a few scattered stations exceed 1000 feet in altitude, and the average height is only a few hundred feet.

GRAVITY VALUES AND THEIR INTERPRETATION

DISTRIBUTION OF STATIONS

The initial attack was made on the Connecticut Valley Triassic unit, and the network of pendulum stations is closest in that part of the region (except for the small area around Providence, studied independently by C. W. Brown). At best, however, the number of pendulum stations is sharply limited by the time required for their occupation; the average rate in this project was one station per day. The most effective arrangement of stations was judged to be in east-west lines, across the structural grain. Four principal lines cross the Triassic belt and include stations on the older rocks at each end; from south to north these lines are designated the New Haven, Meriden, Hartford, and Springfield profiles. The closest spacing is in the New Haven profile, where the average interval between stations is about 2 miles. For the other profiles the average is about 5 miles. Supplementary stations outside the profiles are located chiefly on both sides of the great fault along the east boundary; a few are similarly grouped along the west boundary.

Three of the profiles were later extended to give a better regional setting. The Meriden line was projected east and west to reach across Connecticut with an average interval of about 6 miles; the line was oriented to cross the Pomperaug Valley Triassic area and the Preston gabbro mass. A few more widely spaced stations connect the Hartford

profile with stations in the Providence network. The Springfield profile was extended eastward toward Plymouth, and westward, with an average interval of 25 miles, into the Catskills. To spread the network northward, a new line of stations, spaced on the average 12 miles apart, was run from southern New Hampshire to the southern part of the Adirondacks.

The gravimeter stations later occupied by Woollard were located to fill critical gaps in various parts of the map, and in particular to give adequate data in the vicinity of the Highlands of the Hudson, on Long Island, and in eastern Massachusetts. It is hoped that eventually isostatic values will be available for all the new stations, to permit accurate comparative study of the several types of anomalies in the entire region.

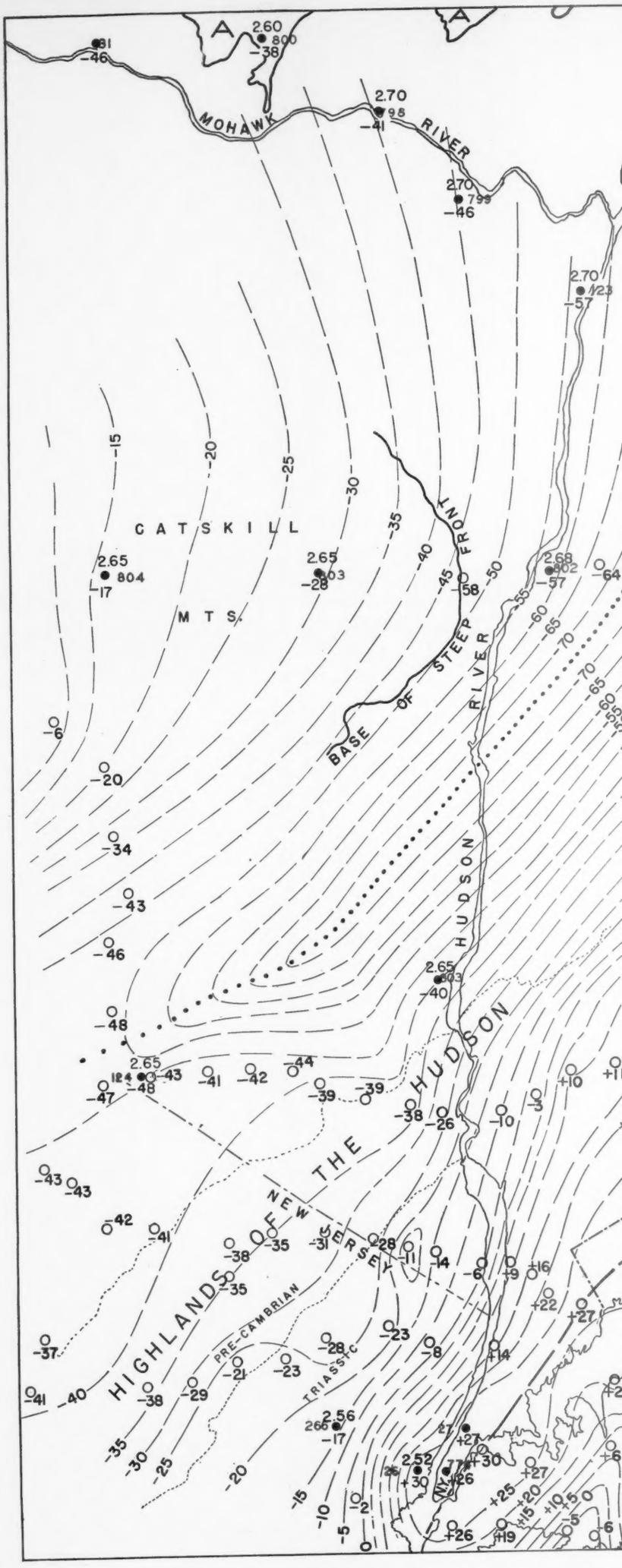
COMPARISON OF ISANOMALY MAPS

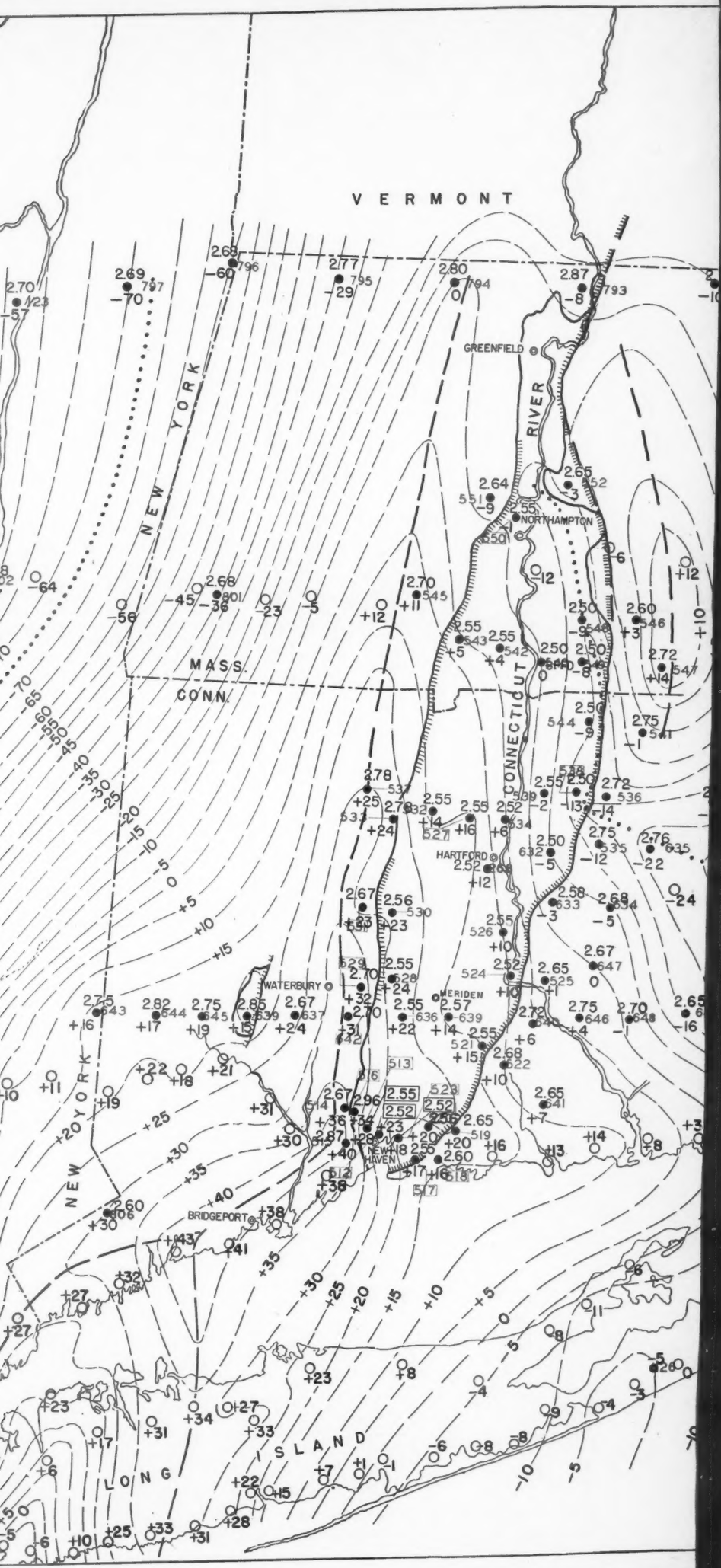
For the isostatic isanomaly map (Pl. 1), anomalies based on assumed compensation at depth 56.9 km. are used, because these values seem most appropriate for this low-altitude region.[1] The writer has prepared another map using anomalies computed on compensation at depth 113.7 km., and the two maps differ surprisingly little. The effect of reducing the assumed depth of compensation is to increase the anomalies algebraically, making all positive values numerically larger, negative values numerically smaller. Figure 7 shows anomaly profiles for the two assumed depths. A profile based on the 96 km. depth would lie in the space between the two shown in Figure 7. Isanomaly maps drawn for the three depths show individual lines relatively displaced but patterns that are strikingly similar, even in detail.

It is even more striking that the isostatic, Bouguer (Pl. 2), and free-air (Pl. 3) maps agree remarkably in their major features.[2] The free-air map is unique in some respects; it reflects some topographic control in the areas of greatest relief, as is to be expected since free-air values involve no correction for the mass above sea level. This topographic influence, in comparison with the isostatic and Bouguer maps, appears on the free-air map in a much steeper anomaly gradient at the Catskill front; in distinct "gravity highs" at stations 643 and 795, which are on high ridges; and in numerous smaller features that clearly are related to land relief. The assumptions on which the free-air reduction is

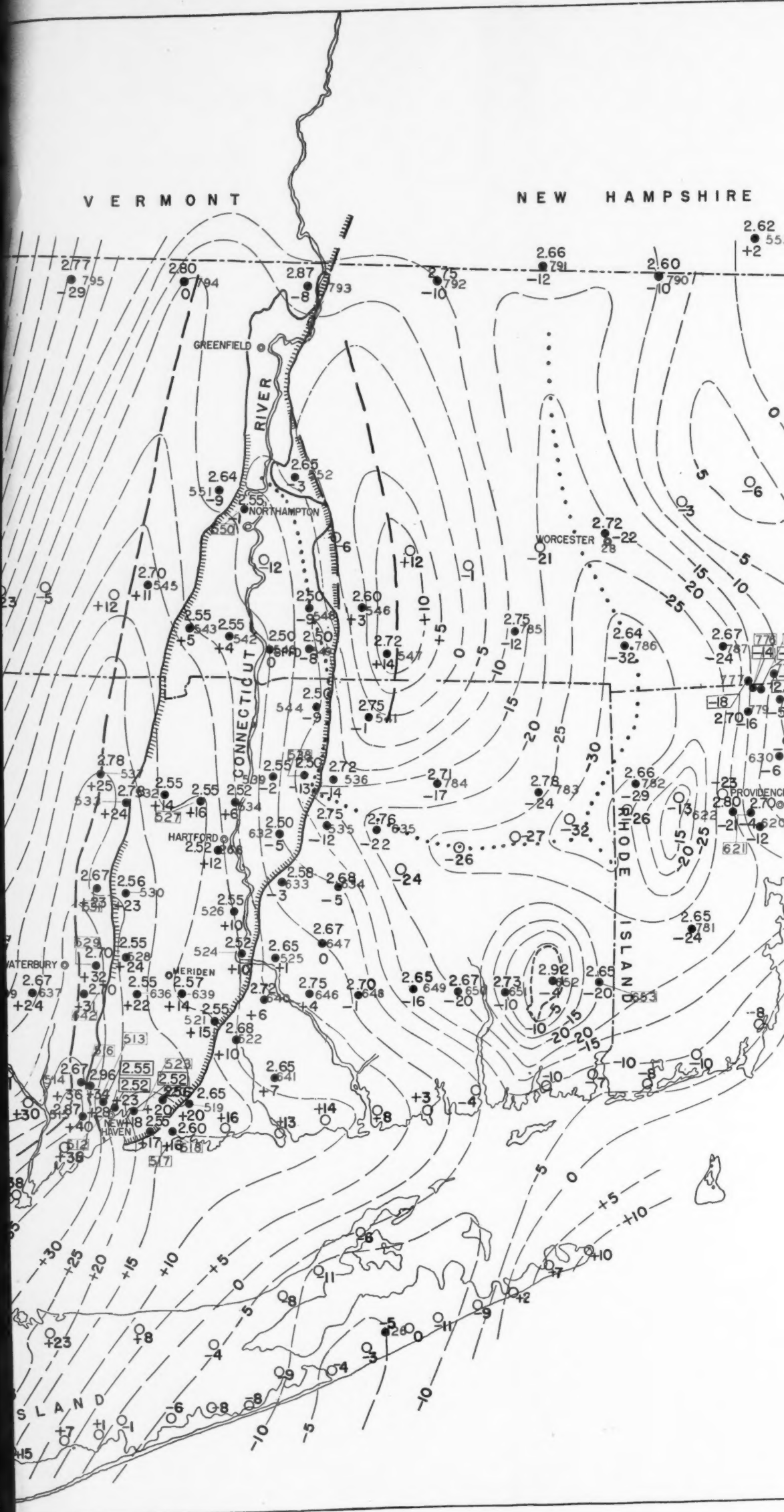
[1] Bowie (1924) found that for mountain stations considered alone, the most satisfactory results are given by assuming 96 km. as the depth of compensation; that for all stations taken together, the best results are given for an assumed depth of about 60 km. On the basis of Airy's hypothesis of a light outer shell with variable thickness, the depth of compensation in a region of low elevation is much less than in a mountain district.

[2] The agreement is more striking when only the pendulum stations are used as the basis for all the charts. Presumably the use of gravimeter values in Plates 1 and 3 would increase their similarity to Plate 2.

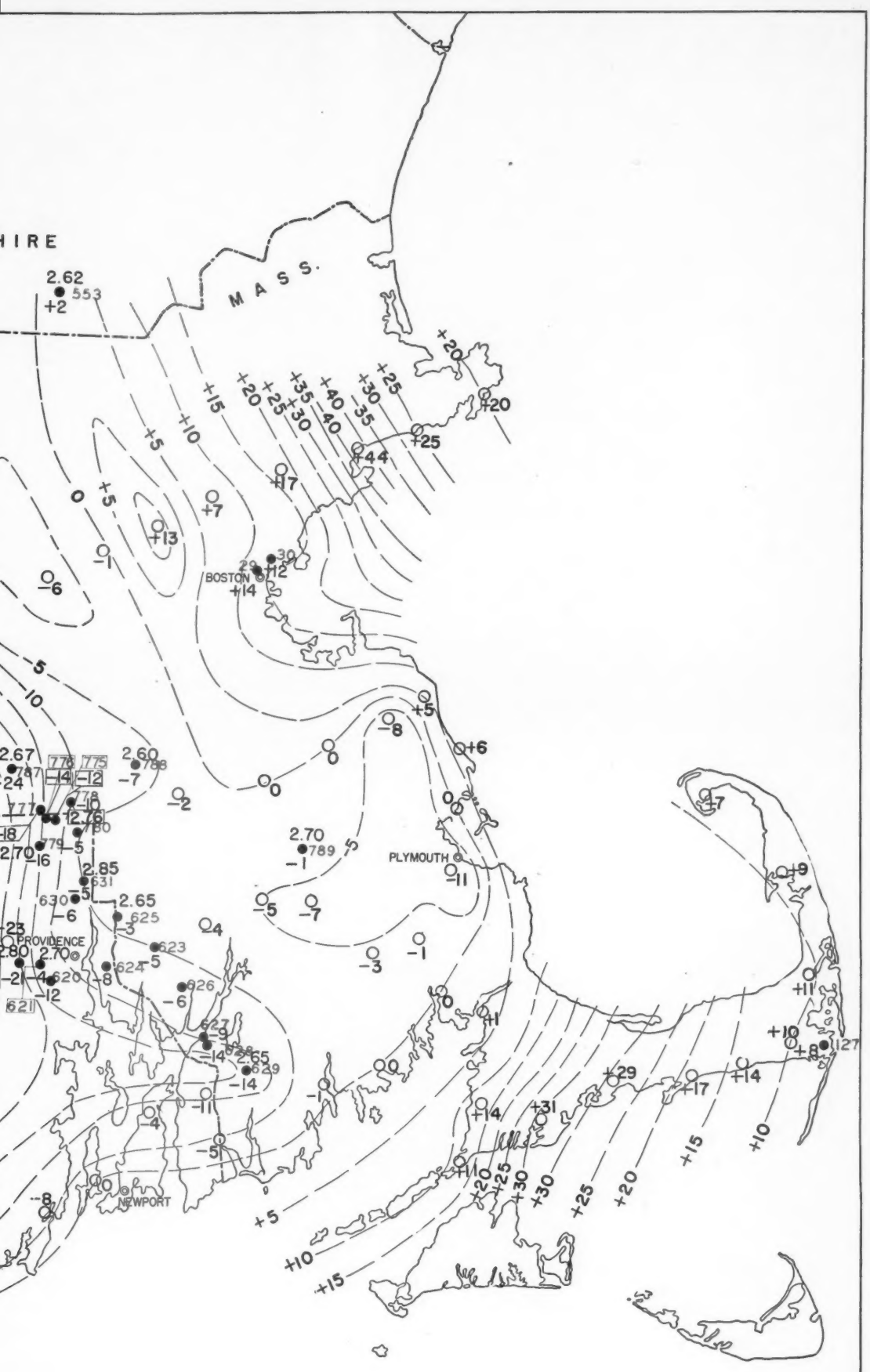

MOHAWK RIVER
CATSKILL MTS.
BASE OF STEEP FRONT
HUDSON RIVER
HUDSON
THE HIGHLANDS OF NEW JERSEY
PRE-CAMBRIAN
TRIASSIC
A
A'
2.60
2.70
2.65
2.68
2.56
2.52



ISANOMALY MAP, BASED ON BOUGUER



ISANOMALY MAP, BASED ON BOUGUER ANOMALIES



ISANOMALY MAP, BASED ON BOUGUER ANOMALIES. FIGURES SHOW ANOMALIES IN MILLIGALS, AND AVERAGE SPECIFIC GRAVITIES OF ROCKS NEAR STATIONS.
STATION NUMBERS IN RED (PENDULUM STATIONS ONLY).
SPECIAL SYMBOLS EXPLAINED ON PLATE I.
○=GRAVIMETER STATION; ● = PENDULUM STATION.

based are highly improbable, and the anomalies are given little emphasis in the present study. Inclusion of the map seems justified because it illustrates so well the weakness of the free-air method and because, in spite of this weakness, it agrees well with the other two maps in the major features depicted. This agreement seems to have special significance, as discussed in a later paragraph.

The outstanding features brought out by the three maps are the following:

(1) A pronounced belt of negative anomalies in the Hudson Valley region. Distribution of these anomalies with like sign suggests a deep "gravity trough," with its axis trending northeastward, its sides sloping up rather uniformly east and west from the axis. The isostatic and Bouguer maps of the "trough" differ somewhat in position of the axis, in depth of the "trough," and in form of cross profile, the Bouguer showing marked asymmetry, with the gentler slope on the west side. Bouguer anomalies for Woollard's line of stations south of the Catskill Mountains indicate that the axis of the "trough" rises sharply toward the southwest. It is to be expected that additional stations in other parts of the region will indicate other modifications in the form of the "trough," especially in a belt near the New York-Connecticut boundary, where the geology is complex. Woollard's Bouguer values in the Springfield-Catskill profile fit remarkably well into the isanomaly lines drawn by interpolation between widely spaced pendulum stations. His Bouguer anomalies east of the Hudson in his southern profiles also check reasonably well. His free-air values in both these areas agree less well with the interpolated lines. (Pl. 3.)

(2) A north-south positive belt in western Massachusetts and Connecticut. The inferred position of the axis of this "gravity high" is closely similar in the three maps. All agree also in depicting a small "gravity plateau" in southwestern Connecticut. Woollard's anomalies in that area suggest numerous small irregularities on the surface of this "plateau." The gravimeter values indicate a well-defined fork in the high axis, one branch extending southward across Long Island, the other reaching southwest to the vicinity of New York City.

(3) A group of anomalies, consistently negative, that appear to outline a "gravity trough," with its axis extending across east-central Massachusetts southward into Rhode Island and eastern Connecticut. This "low" is less pronounced and more irregular in outline than the analogous feature in the Hudson Valley region. From what appears to be its northern extremity, near the north boundary of Massachusetts, it widens and "pitches" southward until it is joined by a branch that extends

eastward from central Connecticut. Woollard's data suggest that the "trough" may terminate southward against a positive axis offshore (Pl. 2). The gravimeter values in the profile northwest of Boston agree reasonably well with the interpolations from scattered pendulum stations. However, the new anomalies suggest local irregularities that will require more stations to delimit. Woollard's Bouguer values indicate that the axis of the Rhode Island "trough" lies somewhat farther west than had been assumed. (*Compare* Plates 1 and 2.)

In addition to these three major features, the maps indicate a "high" directly east of the Triassic area in Massachusetts and northern Connecticut, and a strong local "high" in eastern Connecticut, on the Preston gabbro mass. The north-south extent of the Preston gravity feature is assumed on the maps, since only the east-west profile is available. Probably the most surprising fact that emerges from study of the maps is their failure to reflect clearly the position of the great fault at the eastern margin of the Triassic area in southern Connecticut. In a considerable stretch east of Springfield, Massachusetts, the fault seems to be strongly indicated in the distribution of anomalies. From north-central Connecticut to Long Island Sound, however, isanomaly lines cross the known position of the fault obliquely, and there is no apparent break in the anomaly gradient anywhere near the fault zone. Similarly, the western margin of the Triassic area in northern New Jersey is not reflected in the anomalies. These unexpected developments from the study are discussed in a later section, in connection with the geology.

Woollard's gravimeter data indicate an important "high" in northeastern Massachusetts and another on Cape Cod (Pl. 2). Isanomaly lines based on these data are of course tentative, since nothing is known of gravity conditions east of the Massachusetts shore line. Possibly the two "highs" coalesce to form a continuous north-south positive axis, analogous to the one in western Connecticut and Massachusetts. Some comparisons between these two belts, as seen on the Bouguer map, have considerable interest. The positive anomalies are closely comparable in size, with a maximum of +40 in the western belt, +44 in the eastern. Spacing of the two positive belts agrees strikingly with the spacing of the two principal negative belts; axes of like sign are about 150 miles apart in the southern part of the region and converge northward in about equal measure. If we consider each "high" belt as paired with the adjacent "low" belt, the axes in the western pair have notable parallelism, with southwest trend, and the axes in the eastern pair agree in a trend somewhat east of south.

RELATION OF ANOMALIES TO THE VISIBLE GEOLOGY

Explanation of specific-gravity values.—Rock samples representing the lithology at nearly all the pendulum stations have been collected and their specific gravities determined [3]—a laborious task, because the geology is highly complex. Even the comparatively simple Triassic sedimentary section presents difficulties, because its beds differ considerably in kind and in density, are lenticular, and have little apparent order in their arrangement. Additional complications arise from the included lavas, sills, and dikes. When all these variable quantities are considered, clearly a consistent rule has to be adopted in arriving at specific-gravity values that are to represent the stations. On the Triassic area, for example, should the attempt be made to calculate an *average* specific gravity for the entire thickness of Triassic rocks, or for all rocks to a standard depth, beneath each station? At first thought such average values seem desirable, as indices of the total geologic effects on anomalies at the several stations. Actually they would lead to confusion in a regional study. For example in Figure 1A, station 543 lies above a comparatively thin sedimentary section. Stations 542, 540, and 549 have progressively thicker sections, each including heavy basaltic lavas. Even if the thickness of the lavas were the same in the three sections, giving the lava uniform value in computing average densities would not yield an accurate indication of the gravitative effect at each station, because of the variable depth of the basalt below the surface.[4] If this complication were adjusted in computing the averages, the resulting values still would not suggest total geologic effects; the column at 543, which would have the lowest density so far as the Triassic section is concerned, probably has the highest average density computed to some standard depth below the Triassic wedge, since it has the largest volume of dense basement rocks.

This Triassic section is used as an example, because its elements can be analyzed with some success. At stations on the metamorphic and igneous complex that underlies most of the region, it is not possible to say how deep any rock unit extends or what form it has below the

[3] This has not been done for a few of the older stations near the borders of the region—*e.g.*, Nos. 29 and 30 near Boston, No. 127 on Cape Cod, No. 126 on Long Island, and No. 131 at the extreme northwest corner of the map—because some of these stations are not in critical positions, and a thick cover of unconsolidated sediments at some of the locations makes it impossible to secure representative bedrock samples. Only a few stations in the network near Providence were selected for specific gravity determinations, since a refined study of a small area within a region whose stations are generally far apart would have little significance from the viewpoint of the present investigation.

[4] Although the lavas have large lateral extent compared with their distance from most of the stations, they have not been treated as plane plates of uniform density and infinite extent, for the following reasons: In parts of the area they are steeply inclined through considerable distances, and in southern Connecticut their lateral extent is small. Moreover, careful study of the anomalies suggests strongly that the gravitative effect of the inclined sheets varies considerably with distances from stations.

surface. Hence, if the specific gravities shown on the maps are to have approximately uniform significance, they must represent the rocks exposed around the stations. Use of this rule makes it possible to compare the effects of visible masses throughout the map area. Wherever there

TABLE 1.—*Examples of specific-gravity values*

1. TRIASSIC SEDIMENTARY ROCKS

Station No.	Specimen No.	Lithology	Sp. Gr.	Av. Sp. Gr.
532	1 & 2	Fine-grained red ss	2.48	
	3	Fine-grained red ss	2.42	
	4	Micaceous siltstone	2.60	
	5	Compact red siltstone	2.66	2.52
540	1	Tuffaceous ss	2.21	
	2	Tuffaceous ss	2.26	
	3	Medium-grained red ss	2.50	
	4	Fine-grained red ss	2.57	
	5	Compact red siltstone	2.66	
	6	Porous fine-grained ss	2.33	2.50
639	1	Coarse-grained red arkosic ss	2.56	
	2	Medium-grained red arkosic ss	2.53	
	3	Compact red siltstone	2.63	
	4	Fine-grained red ss	2.59	
	5	Medium-grained red ss	2.55	2.57

2. GRANITIC ROCKS

Station No.	Specimen No.	Lithology	Sp. Gr.	Av. Sp. Gr.
522		Haddam granodiorite gneiss.		
	1	Roof pendant(?) of dark biotite gneiss	2.76	
	2	Medium-grained dark-gray diorite	2.74	
	3	Coarse pegmatite (graphic granite)	2.57	
	4	Fine-grained gray granite, faintly banded	2.60	2.68
551		Williamsburg granodiorite.		
	1	Coarse-grained, gray, massive, rich in biotite	2.65	
	2	Coarse-grained, gray, massive, rich in biotite	2.64	
	3	Finer-grained and lighter-colored than (1) and (2), with less biotite	2.62	
	4	Finer-grained and lighter-colored than (1) and (2), with less biotite	2.60	2.64
782		Sterling granite gneiss.		
	1 & 2	Strongly banded, rich in biotite, has scattered garnets	2.67	
	3	Similar to (1) and (2), but coarser-grained	2.63	
	4	Similar to (3)	2.64	2.66

3. METAMORPHIC ROCKS

Station No.	Specimen No.	Lithology	Sp. Gr.	Av. Sp. Gr.
533		Collinsville gneiss.		
	1	Black hornblende gneiss	3.07	
	2	Garnetiferous mica gneiss	2.87	
	3	Mica gneiss, with *lit-par-lit* pegmatite	2.70	
	4	Mica gneiss, with *lit-par-lit* pegmatite	2.73	
	5	Dark biotite gneiss	2.90	2.78
643		Hartland(?) schist.		
	1	Coarse-grained biotite schist with scattered garnets	2.80	
	2	Similar to (1), with more abundant garnet	2.92	
	3	Biotite schist, no visible garnets	2.71	
	4	Biotite schist, no visible garnets	2.70	2.76
795		Rowe and Hoosac schists.		
	1	Lustrous phyllite (Rowe)	2.89	
	2 & 3	Lustrous greenish schist (Rowe)	2.82	
	4	Albitized biotite schist, coarse-grained (Hoosac)	2.68	
	5	Albitized biotite schist, coarse-grained (Hoosac)	2.73	
	6	Albitized biotite schist, coarse-grained (Hoosac)	2.80	2.77

4. PALEOZOIC MARINE FORMATIONS IN HUDSON VALLEY

Station No.	Specimen No.	Lithology	Sp. Gr.	Av. Sp. Gr.
798		Trenton(?) formation.		
	1	Impure limestone, with partings and lenses of gray shale	2.67	
	2	Similar to (1)	2.65	
	3	Impure gray dolomite	2.79	
	4	Similar to (3)	2.78	2.70
805		"Hudson River" shale.		
	1 & 2	Impure dark-gray dolomite	2.78	
	3 & 4	Dark-gray clay slate	2.53	
	5	Slaty argillite, gritty	2.67	
	6	Dark gray to black argillite	2.71	2.65

is some information on the subsurface geology, as in the Triassic belt, effects of the several elements can be computed as a special problem.

Samples taken from chance outcrops near stations may be misleading. The practice followed in collecting specific-gravity samples was to survey around each station an area with a radius of half a mile or more and to collect specimens from every unit that appears to form a considerable part of the bedrock. Not uncommonly as many as 10 or 12 specimens were taken for a station. This practice is especially advisable for locations on the metamorphic rocks, which consist of diverse types of gneiss, schist, and

phyllite, injected in varying degree with granite and other plutonic rocks. Such a complex poses a serious problem not only in choosing representative samples, but also in assigning to each sample its proportional part in

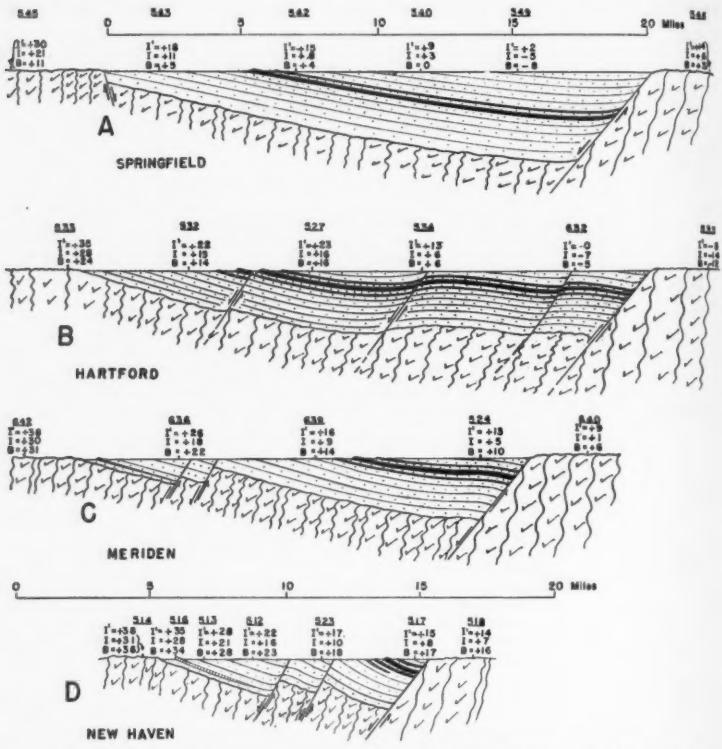


FIGURE 1.—*Generalized east-west cross sections through the Triassic belt of Massachusetts and Connecticut*

A, section through Springfield, Mass.; B, section north of Hartford, Conn.; C, section 2 miles south of Meriden, Conn.; D, section near New Haven, Conn. Thickness of lava flows (black) somewhat exaggerated. In some sections the flows are more deformed, especially by faulting, than drawings show.

Numbers of stations that lie approximately on lines of the sections are shown at top of each (underlined). $I' =$ isostatic anomaly assuming depth of compensation 56.9 km.; $I =$ isostatic anomaly assuming depth 113.7 km.; $B =$ Bouguer anomaly.

Dolerite sills shown by cross lining; pre-Triassic rocks represented by wavy lines and check marks. (Approximate locations of sections shown by lines on Fig. 3.)

the composite specific-gravity value. Table 1 illustrates the range in densities of samples at a given station and the degree of personal judgment required in computing averages.

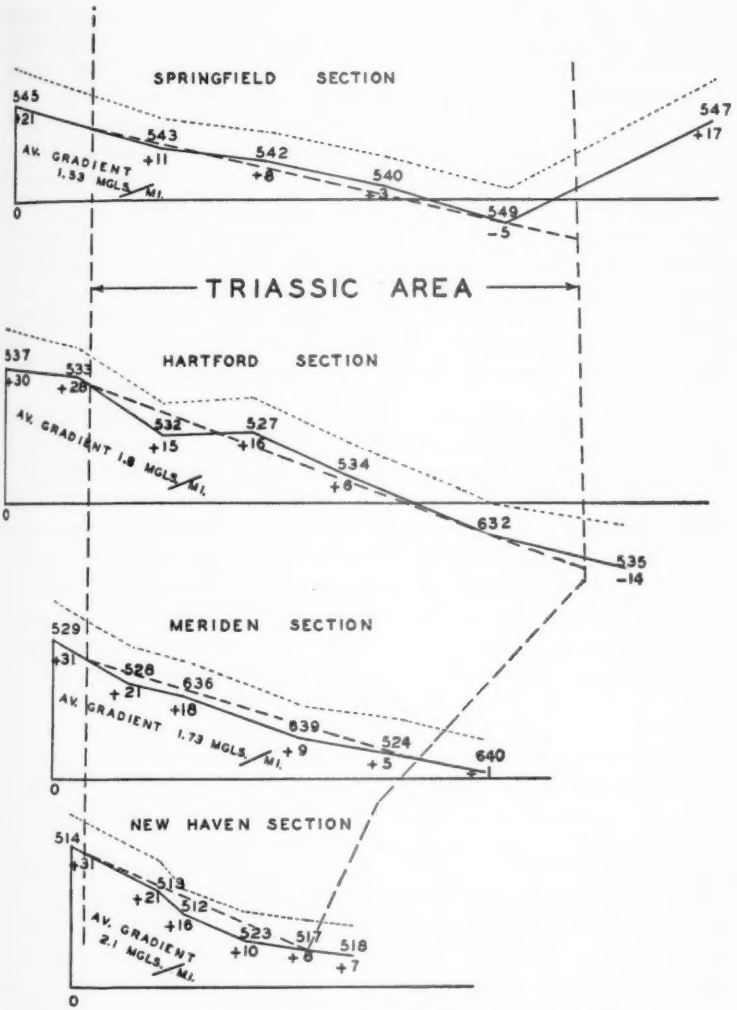


FIGURE 2.— *Anomaly profiles corresponding to cross sections of Figure 1*

The "average gradients" represent the slopes of dash lines drawn from western boundary of Triassic area to easternmost Triassic stations in the several sections. Profiles in solid lines drawn on basis of isostatic anomalies, depth of compensation 113.7 km.; anomalies shown below. Dotted profiles are for depth of compensation 56.9 km. (values of anomalies omitted, to avoid confusion of numbers).

The specific gravities were measured with a specially constructed balance equipped to handle large hand specimens. Several clear quartz crystals were used to check the accuracy in performance and calibration of the balance; all the crystals gave the value 2.65, with minor differences in the third decimal place. Specific gravities of the rock specimens were determined to the second decimal place only, since the limits of error in several elements of the problem do not warrant greater refinement. In preparation for measurement the samples were immersed in water for several weeks, and the surface of each sample was wiped with a cloth immediately before it was weighed in air. It is believed that results obtained in this way represent the densities of the rock masses in their underground occurrence as satisfactorily as would values given by the more laborious Russell-tube method (Chamberlin, 1934), in view of the several possible sources of error in the present problem.

Connecticut-Massachusetts Triassic belt.—Figure 1 represents somewhat generalized east-west geologic sections across the large Triassic area near Springfield, through Hartford, a short distance south of Meriden, and near New Haven.[5] Because each section contains a wedge of light-weight sedimentary rocks, thickest near the east end and feathering westward, it is to be expected that gravity anomalies distributed along the sections will decrease algebraically from west to east. Isostatic and Bouguer anomalies, shown with the sections, conform to this general expectation. It would be expectable, also, for abruptly higher anomaly values to appear at the east ends of the sections, on the pre-Triassic rocks in the foot-wall block of the great fault. The expectation is met in the Springfield section only; in sections farther south the decrease in anomaly values continues across the position of the fault.

The anomaly profiles corresponding to the four sections are plotted in Figure 2. A break in each profile does occur near the fault, although in the three profiles south of Springfield the direction of slope is not reversed. In each of the three profiles a straight line drawn from the western edge of the Triassic area through the position of the easternmost station in the area has a gradient distinctly steeper than that of the profile continued east of the fault. For the Hartford section this straight line happens to have a position near the average for all the Triassic stations concerned. If such an "average profile" were drawn for the Meriden and New Haven sections also, the break in profile near the eastern boundary would become more distinct in each of these sections. Thus the local geology appears to be reflected in all the profiles, but most clearly in that

[5] The sections are constructed from surface evidence. A seismic project by D. Linehan, begun in 1941, may yield more accurate information on the thickness of the Triassic wedge.

for the Springfield section. Since the great fault has large throw in southern Connecticut, the faint reflection in the gravity profiles calls for special explanation.

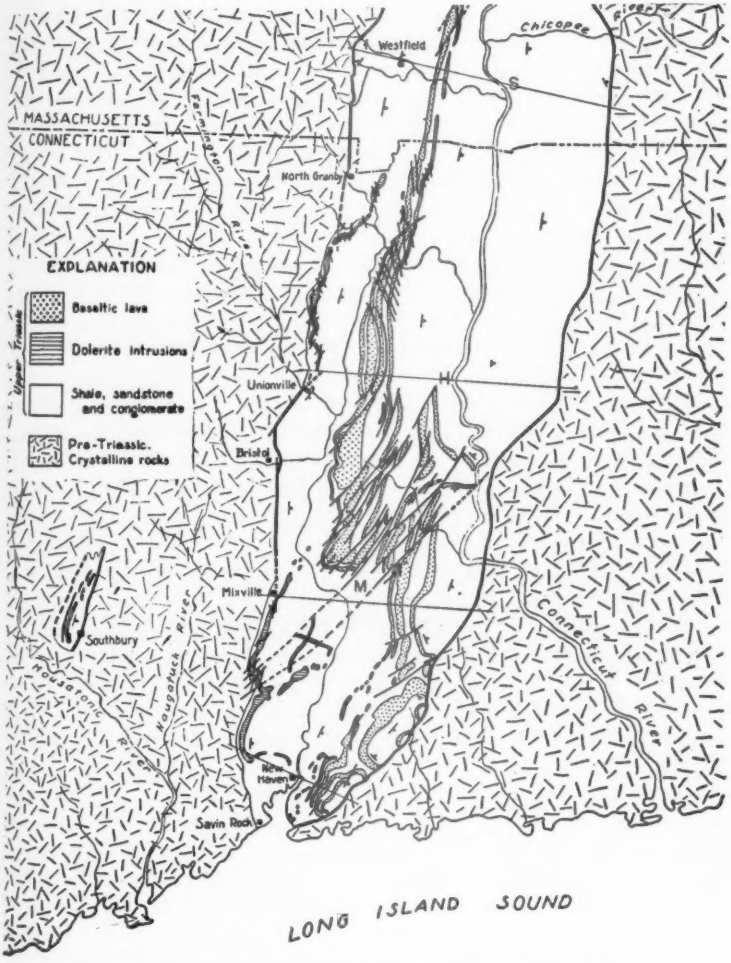


FIGURE 3.—*Map showing basaltic lavas in the Triassic area*

Connecticut and southern Massachusetts. Locations of Springfield, Hartford, Meriden shown by *S*, *H*, *M* respectively. Lines crossing Triassic area indicate approximate locations of sections in Figure 1. (After Wheeler, 1937.)

A cause is suggested in the distribution of basaltic lavas, as shown in the structure sections (Fig. 1) and on the geologic map (Fig. 3). Several faults cross the Triassic area obliquely, from northeast to southwest. Downthrow was consistently on the northwest, and hence the blocks are progressively higher, in a structural sense, southward. Erosion planed the blocks to near a common level; hence outcrops of the eastward-dipping lavas are offset eastward toward the south, and near the eastern boundary fault the buried flows are at progressively shallower depths southward. In southern Massachusetts and northern Connecticut, stations on the Triassic near the great fault are on a large thickness of light-weight sedimentary rock, which influences gravity values much more than the deeply buried lavas. In southern Connecticut the lavas, because of their nearness to the stations, more nearly offset the effect of the sedimentary section, which moreover has a smaller total thickness than farther north. Sills and dikes are intruded only below the highest lava, and hence these intrusive bodies also lie closer to the eastern stations in the southern part of the area. It may be urged that the algebraic increase of anomalies by heavy rock near the eastern stations should tend to lower the average anomaly gradient in the southern Triassic sections, whereas the gradient actually is highest in the New Haven section. However, the greater regional gradient at New Haven, as compared with Springfield, is sufficient to overcome the local effect within the Triassic belt. This point becomes clearer after the following analysis.

Approximate effects of the Triassic sediments and basaltic flows were computed, and corrections were made in profiles for the four Triassic sections, as shown in Figure 4. Bowie's method and table of values were used (1924, p. 37-38).[6] Exhaustive computations were not attempted. Although all parts of the Triassic prism affect the gravity value at every station in the area, the actual effect beyond a radius of a few miles from a station is slight, because of the small depth to which the prism extends. For the calculations an average specific gravity of 2.55 was assumed for the Triassic sedimentary rocks, 2.97 for the basalt. Thus in comparison with the density 2.67 assumed in computing anomalies the sedimentary part of the section has a defect of 0.12, the basalt an excess of 0.30. The net corrections, additive for all stations because the basalt is at best the minor element, tend to make the profiles more regular and to decrease the average slopes (Figure 4). The corrected profiles may be taken to represent the regional gradients, freed from the local effects of the Triassic wedge. This gradient is least in the Springfield section, progressively larger in the Meriden and New Haven sections, and

[6] Approximate corrections are computed quickly by the following formula: 1000 feet of rock whose excess or defect in density is 1 creates a gravity anomaly of 13 milligals. However, this rough formula ignores the lateral extent of the disturbing mass, which is taken into account by Bowie's method.

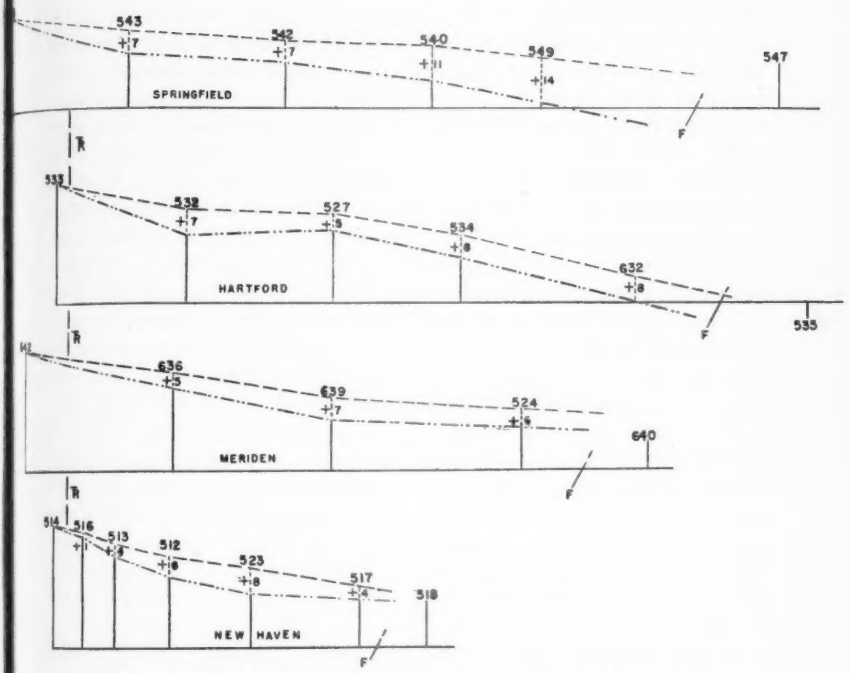


FIGURE 4.—*Correction profiles (upper dash line) for abnormal densities of the Triassic rocks*

Obtained by adjusting isostatic anomaly profile (dash-and-dot line) for the effects of low-density strata and high-density basalt near stations. Net correction at each station indicated by short-dash vertical line, with numerical value and sign as shown. Anomalies used are for depth of compensation 56.9 km. (*Compare* Figures 1 and 3 for differences in local geology among the four sections.) F = eastern boundary fault.

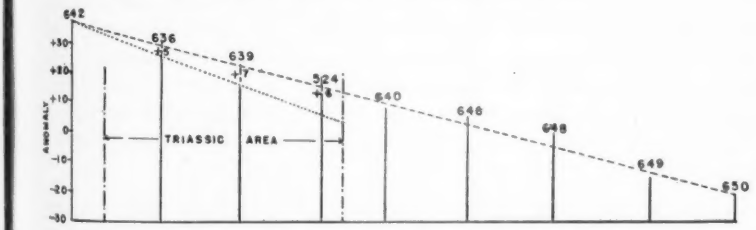


FIGURE 5.—*Meriden anomaly profile extended eastward on the older rocks*

Dotted line shows average gradient on Triassic area, neglecting station near great fault. Corrections for abnormal densities indicated by vertical dash lines. Upper dash line shows average gradient for entire line of stations. (Anomalies for depth 56.9 km. used; base of diagram at —30 milligals.)

greatest in the eastern part of the Hartford section—evidently from the influence of the "gravity syncline" in eastern Connecticut, which does not affect the other Triassic sections. The correction diagrams present several points of special interest. In the Springfield section the space between uncorrected and corrected profiles is widest and continues to widen eastward—an indication of the sedimentary wedge as the preponderant local influence. Reduced eastward divergence of the curves for the Hartford section and actual convergence in the Meriden and New Haven sections testify to increasing effect of the basalt southward. Failure of the correction entirely to remove the rise in the profile at station 527, in the Hartford section, suggests that the basalt may be at shallower depth under the station than the structure section assumes, or that a hidden intrusive mass underlies the station. Possibilities of incompleteness in the corrections are obvious in all the profiles. The corrected curve extended eastward from station 549 falls below the anomaly value at station 547, perhaps because the thickness of Triassic deposits or their defect in density is larger than assumed for that area.[7] In the Meriden and New Haven sections the corrected profiles, projected eastward, pass above the stations on the older rocks. However, the values for these stations also require small corrections for the effect of the Triassic wedge. Moreover, the effect of intrusive bodies in the southern sections, not possible to evaluate quantitatively, would be to lower the corrected profiles somewhat for the stations on the Triassic.

Further analysis is given in Figure 5 for the Meriden profile extended eastward to include five stations on the pre-Triassic rocks. An average anomaly profile across the Triassic is assumed to be a line (dotted) that fits stations 636, 639, and 642—the latter on the older rocks near the western edge of the Triassic. Stations on the Triassic near the eastern border (either 521 or 524) have values well above the average profile. There are two logical causes of this irregularity: Both the lavas in the Triassic section and the heavy rocks east of the great fault tend to offset the defect in mass of the Triassic sediments. Corrected values for the three Triassic stations lie near a straight line drawn between stations 642 and 650, at the ends of the profile. This line is also a fair average gradient for the five stations east of the Triassic belt; individual variations from the average are not explained by anything known in the complex local geology.

Effects of other exceptional rock masses.—Figure 6 presents a profile for a continuous line of stations extending across Connecticut from eastern New York to near the Rhode Island line. Corresponding specific gravi-

[7] *See* later section on crustal warping for another suggested explanation of the position of the corrected curve at Springfield.

FIGURE 6.—*Anomaly and specific-gravity profiles across Connecticut*
Near Meriden. Outcrop boundaries of principal geologic units are indicated.

ties are included in a parallel profile, and the principal surficial geologic units are delimited. Aside from the large Triassic belt, the units of chief interest in relation to the profile are the small Pomperaug Triassic area and the Preston gabbro mass. In a general way there is good correspondence in form between the two curves. The Pomperaug area appears as a striking exception. However, the one station in that small area is on a thick basalt flow, which has the high specific gravity shown in the profile. When the low density of the associated Triassic sediments is considered, the dip in the anomaly curve at this point is explained qualitatively. However, the correction for the Triassic prism, with its small thickness and areal extent, amounts to less than one-fourth of the drop in anomaly value, as compared with the stations on either side. Some other factor, not detectable in the surficial geology, must be responsible for the dip in the profile. This example illustrates the danger of accepting local geology to explain anomalies without an attempt to make a quantitative check of the effect to be expected.

The Preston gabbro profile excellently illustrates strong reflection of local geology in the anomalies, although it is not possible to check the quantitative effect on the basis of present knowledge. Loughlin (1912) gave reasons for supposing that the main mass of gabbro is laccolithic. A pluglike form is not excluded by the field evidence. If we take the average radius of the horizontal plan as 4 miles and assume the average excess in density as 0.30, Bowie's tabular values (1924, p. 37) show that the mass must have an average thickness of 4000 feet to produce the gravitative effect indicated by the east-west profile. This rough check favors Loughlin's suggestion that the thickness is small in comparison with the lateral dimensions. If the gabbro formed a cylindrical mass reaching to a depth of several miles, its disturbance of the anomaly profile should be several times stronger than it is.

STUDY OF REGIONAL PROFILES

Anomaly profiles have been constructed along two east-west lines, almost across the entire map area; one line from the Catskills through Springfield, to a point in eastern Massachusetts; the other crossing the lower part of the Hudson Valley, southern Connecticut, and Rhode Island. Figure 7 shows the profiles on a basis of isostatic anomalies, representing two depths of compensation. Figure 8 shows profiles for the same two lines of stations, on bases of Bouguer and free-air values. These four sets of profiles agree strikingly in their major features. All indicate a deep "gravity syncline" in the Hudson valley region, a pronounced "high" in Connecticut and Massachusetts, and a Rhode Island-Massachusetts "syncline." All give essentially the same expression to the Preston

gabbro mass and the large Triassic belt. All agree even in many minor details, such as the Pomperaug "dimple" and other moderate irregularities. The free-air curves have peculiar features that result from lack of topographic correction. The Bouguer profiles represent the Hudson Valley "trough" as more asymmetric and deeper than do the isostatic curves, and there are other differences of emphasis.

General parallelism of the Bouguer and isostatic profiles points strongly to a major effect that has no connection with ordinary assumptions of isostatic compensation. To check the obvious suggestion that the explanation may lie in the local geology, specific-gravity curves corresponding to the anomaly profiles are plotted in Figures 7 and 8. A close relation between anomalies and rock densities is clear in some details of the profiles. The most conspicuous responses to density variations are seen in the large Triassic belt and in the area of the Preston gabbro. Other minor effects of local geology are detectable in the curves. However, the irregularities in anomaly profiles that obviously reflect known density differences are superposed on much larger irregularities which appear to be independent of the visible geology. The Preston gabbro, one of the densest rocks in the region, causes a sharp rise in the anomaly profile, but does not suffice, even where the mass is thickest, to overcome the regional negative tendency within the Rhode Island basin. Similarly the Triassic sedimentary section, with average specific gravity much lower than that of any other thick rock unit in the region, causes a "gravity trough" that lies high on the regional curve; the average anomaly on the Triassic belt is rather strongly positive. Finally, the Hudson Valley "gravity trough" finds no adequate explanation in densities of visible rock masses. The shales and carbonate rocks within this belt are nearly as high in average density as the metamorphic and plutonic rocks farther east, and there are no detectable differences in density that correspond to the large variations in negative anomaly values within the "trough" itself.

The regional tendency of anomalies to disregard the local geology is brought out forcibly by Table 2 showing average values in Connecticut, divided into three unit areas from west to east.

The average values emphasize the contrast in gravity conditions between western and eastern Connecticut, although densities of visible rock masses in the two areas differ but slightly. The Triassic belt connects the western and eastern areas logically in a gravimetric sense, although it appears as a sharp hiatus from the viewpoint of surficial geology. Study of the anomaly charts (Pls. 1, 2) shows that the contrast between the western and the eastern areas would be increased if the Rhode Island stations were included. An enlarged table to bring out the principal gravimetric features for the entire region might recognize four units, as

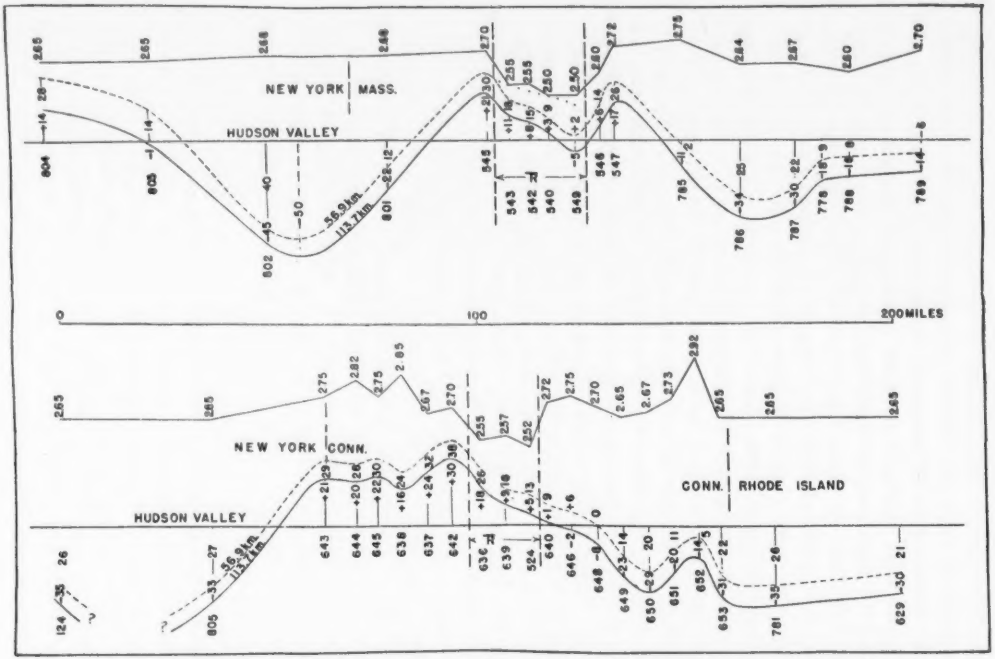


FIGURE 7.—*Anomaly and specific-gravity profiles, eastern New York to eastern Massachusetts*

Profiles in upper half of figure are for line of stations passing through Springfield, Mass.; those in lower half are for line passing near Meriden, Conn. Isostatic anomalies are used, for the two depths of compensation indicated. Station numbers shown below profiles; anomaly values for profiles on each line are shown in positions corresponding to positions of profiles. $\mathcal{T}_R$ = Triassic area. Line of large dots in upper profile, crossing Triassic area, represents correction for abnormal rock densities, as applied to anomalies for depth of compensation 56.9 km.

shown below profiles; anomalies in milligals. $\mathbb{T}$ = Triassic area. Line of large dots in upper profile, crossing Triassic area, represents correction for abnormal rock densities, as applied to anomalies for depth of compensation 58.9 km.

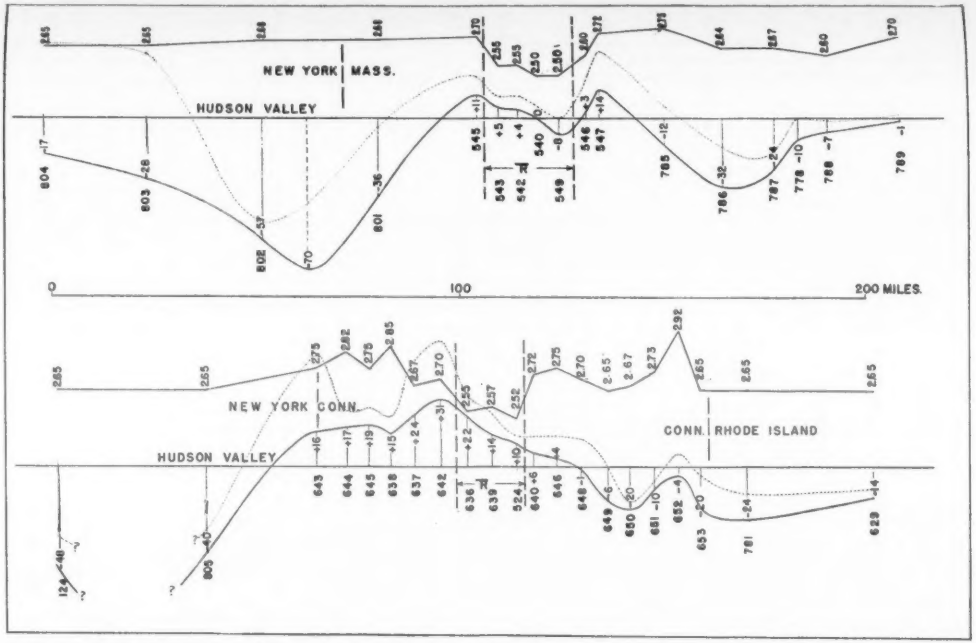


FIGURE 8.—*Bouguer (solid line) and free-air (small dots) anomaly profiles, eastern New York to eastern Massachusetts*

Free-air anomaly values omitted, to avoid confusion. Note sharp rise of free-air profile near Catskill front (upper profile, at left) and large irregularities of free-air profile along the southern line; these features result from omission of topographic correction in the free-air reductions.

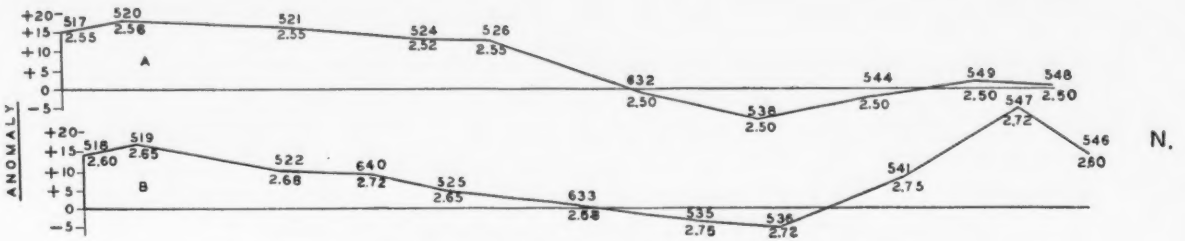


FIGURE 9.—*Anomaly profiles parallel to great fault at eastern boundary of Triassic area*

Station numbers are given, with specific gravity at each station. Upper profile represents stations on Triassic area, near the fault; lower profile represents stations east of the fault, on the pre-Triassic rocks.

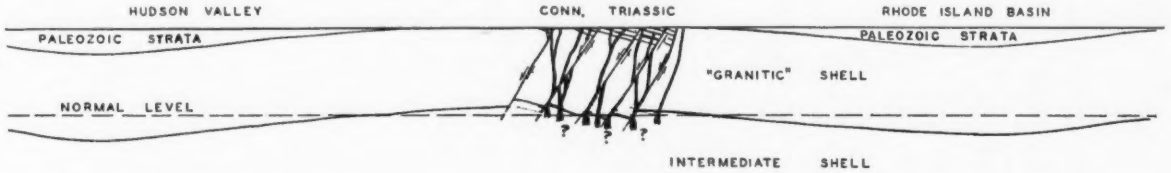


FIGURE 10.—*Hypothetical section showing possible application of crustal-warping hypothesis to New England-Hudson Valley anomalies*

Scale is only approximate, and relations are much simplified. Paleozoic rocks, deformed and metamorphosed, probably extend across middle part of section, but they are assumed to be thin there. Deformation of strata in both basins, and their intrusion by granitic masses, not represented. Mafic rocks in Triassic part of section represented as injected in part along faults. Present surface of low relief indicated by upper straight line.

follows: (1) eastern New York (strongly negative); (2) western Connecticut and Massachusetts (strongly positive); (3) the Triassic belt (less strongly positive); (4) the area east of the Triassic belt (negative).[8] Brief study of the isanomaly maps indicates clearly that average specific gravities for the four units do not vary with the average anomalies.

TABLE 2.—*Pendulum gravity stations in Connecticut*

	Number of stations	Average anomaly (in milligals) with regard to sign		Average specific gravity
West of Triassic Area	12	Isostatic (56.9 km.),	+34.2	2.75
		Isostatic (113.7 km.),	+26.5	
		Bouguer,	+28.0	
In Triassic Area	21	Isostatic (56.9 km.),	+16.6	2.53
		Isostatic (113.7 km.),	+ 9.6	
		Bouguer,	+13.2	
East of Triassic Area	22	Isostatic (56.9 km.),	− 1.4	2.70
		Isostatic (113.7 km.),	−10.3	
		Bouguer,	− 4.8	

The east-Connecticut branch of the Rhode Island "trough" also appears to be independent of the known geology. Reference to the geologic map of Connecticut shows that this feature is essentially at right angles to formation boundaries. Figure 9 shows the expression of the branch "trough" in north-south anomaly profiles parallel to the great fault at the eastern boundary of the Triassic belt. The "trough" extends across the position of the fault and has no logical relation to specific gravities of the visible rocks. Presumably the unknown cause lies at some depth.

POSSIBLE EXPLANATION IN CRUSTAL WARPING

COMPARISON WITH INDIA

Glennie (1932) pointed out that the anomaly values in India are not materially improved by ordinary isostatic reductions and suggested an explanation in large-scale crustal warping. Wherever there has been long-continued uplift, with attendant erosion, dense rocks in the deeper parts of the crust have been brought closer to the earth's surface. Similarly, secular downwarping accompanied by sedimentary accumulation has lowered the upper surface of a high-density shell for which there is geologic

[8] Only pendulum stations are included in the grouping.

and geophysical evidence. In India a wide east-west belt of positive anomalies is flanked by large negative belts. The central area has long been subject to uplift and erosion; the northern belt with defective gravity values is the Himalayan-Gangetic trough, in which a thick sedimentary fill has been further thickened by folding and thrusting. A logical consequence is illustrated in Figure 11. Following H. Jeffreys, Glennie assumes that the crust consists of an outer "granitic" (sial) shell, a lower shell of dunite, and a separating shell of intermediate density. Continued uplift and erosion in the central belt of India have brought the tops of the intermediate and dunite shells higher in the crust, with resulting increase in their effects on gravity as measured at the surface. Continued depression of the deeper shells in the Himalayan-Gangetic region has decreased their influence on measured gravity. The later sediments deposited in front of the Himalaya chains have deficient density, but these deposits are thick enough to account for only a small part of the defect in measured gravity.

Consideration of the geologic history makes the Glennie hypothesis attractive as a possible explanation of the anomaly distribution in eastern New York and southern New England. The two Paleozoic troughs are separated by an axis of uplift (Fig. 10). Sinking of the basin floors while sediments accumulated accounted for only part of the crustal warping; probably a larger part resulted from successive orogenies, which deformed and thickened the sedimentary prisms (and perhaps also the sial shell beneath them). Induration and metamorphism of the sedimentary strata made them essentially as dense as the associated "granitic" shell. Batholithic intrusions into the deformed strata made no essential change in density, since the intruded material was derived from the light outer shell. Thus the specific gravities of surficial rocks differ little from the Hudson Valley to the Rhode Island basin, except for the post-orogenic Triassic deposits and a few local masses of mafic rocks. The assumed arrangement below the "granitic" shell need not agree exactly with that proposed by Jeffreys and Glennie; whether we picture definite shells or merely a gradation to greater density downward, the net effect of the crustal warping must have been to place the denser part of the crust farther from the earth's surface under the orogenic zones, nearer the surface in the uplifted belt between them.

This suggested correlation between anomalies and crustal structure appears highly satisfactory, at least qualitatively, for the section extending from eastern New York to Rhode Island, but less satisfactory for eastern Massachusetts on the basis of present knowledge. The Acadian trough appears to have extended from the Providence region northeastward across the Boston basin. The axis of the Rhode Island "gravity trough"

extends somewhat west of north, and the largest positive anomalies revealed by Woollard's gravimeter survey are in the area northeast of Boston, supposedly along the course of the Acadian trough. However, information for the eastern part of the region is meager in several respects, and there are several possible reasons for the unexpected distribution of anomalies. The location of the Acadian trough is poorly defined, because the early Paleozoic formations have been extensively engulfed by intrusive igneous masses, blurred by metamorphism, and concealed under younger deposits. Not only is the outline of the trough indefinite; there is practically no information as to the measure of subsidence, and therefore almost no basis for making even a rough estimate of the expected effect on gravity values. Whatever may have been the crustal structure related to the geosynclinal sinking, possibly it was modified considerably by orogenic movements. Moreover, there has been no adequate study of the eastern Massachusetts anomalies in relation to the surficial geology. Large masses of dense igneous rocks in the area northeast of Boston may be chiefly responsible for the positive values there. In view of these uncertainties in the basic information, any suggestions based on the data for eastern Massachusetts are preliminary and tentative. The east end of the section in Figure 10 is based entirely on the Rhode Island basin, for which both the geologic and the gravimetric information are reasonably good.

Triassic deposition, faulting, and igneous activity produced a local complication in parts of the upwarped zone in Connecticut and Massachusetts. On the assumption that subsidence beneath the Connecticut Valley Triassic affected the entire thickness of the "granitic" shell (Figure 10), defect of gravity on the Triassic wedge is caused in part by deficient density of the sedimentary deposits, in part by local lowering of the contact between the "granitic" and intermediate shells. In Figure 7, the dotted line across the Triassic area in Massachusetts represents the correction for sub-normal density in the Triassic wedge. Since this correction does not lift the profile to the position of a smooth curve between stations on the older rocks east and west of the Triassic area, it is a logical speculation that the anomalies reflect also some lowering of the base of the "granitic" shell, as suggested in Figure 10. Triassic basaltic rocks are asssumed to have originated in the intermediate shell. The greater influence of these heavy rocks on measured gravity in southern Connecticut, as compared with southern Massachusetts, is reflected in contrast between the two profiles.

The wave length of the inferred crustal warping, as measured between the axes of the Hudson Valley and Rhode Island "gravity troughs," is 100 to 150 miles. There is no geologic evidence to give a measure of the amplitude. By making several fundamental assumptions, such as the depth

to the base of the granitic shell in each part of the region, the thickness of the intermediate shell, and densities of the shells, we could compute geologic corrections to the theoretical gravity values at all stations. Because of the several unknown factors involved, it seems unprofitable to attempt such computations. The reader can appreciate, however, that the effect of corrections based on the "warp" hypothesis would be to lower the high anomaly profile in western Connecticut and Massachusetts and to raise the low profiles in the Hudson Valley and the Rhode Island basin. The degree to which these irregularities in the regional profile were eliminated would depend on values assumed for the unknown factors.

Some geophysical techniques and results already available point the way to possible solution of the problem. According to the "half width" principle (Nettleton, 1940), a departure from a normal gravity profile represents a disturbing mass whose center of gravity is at a depth not greater than half the length of the chord drawn through the middle of the vertical line representing the maximum amplitude. Thus in the curves of Figure 7 the width of each major upbend in the profiles between the Hudson Valley and Rhode Island basin measures 60 to 70 miles along the median chord. By the "half-width" formula the disturbing influence causing the upbend is centered at a depth not in excess of 30 to 35 miles (48 to 56 km.). From analysis of quarry-blast records Slichter (1939) found that at a depth of 23.5 km. beneath the Connecticut Valley the speed of the compressional wave increased from 6.32 km. to 7.82 km. per second. Leet (1938), on the basis of seismic investigations, cited evidence of a discontinuity in New England at a depth of about 14.5 km. Possibly these results represent variations in depth of the base of the "granitic" layer. Future extension of seismic studies may reveal whether variations in depth of an important discontinuity have a systematic relation to the major gravity "highs" and "lows," and may yield dependable data on which to base quantitative corrections for crustal warping.

White (1923, p. 231-232) recognized that the Paleozoic sedimentary rocks of the Appalachian geosyncline are not sufficiently low in density to explain the consistently large negative anomalies throughout the Valley-and-Ridge province. His suggested explanation is that pre-Cambrian rocks directly below the folded Paleozoic section are less dense than the rocks exposed in the adjoining Piedmont belt. This suggestion does not bear analysis. Negative anomalies cannot result from rock densities that are *relatively* low; the rocks beneath and near a station cannot be suspected as the cause of a negative anomaly unless they have average density below the standard value assumed in the computations. The average density of Paleozoic formations in the Appalachian geosyncline appears

to be fully up to this assumed standard, and it is highly improbable that any pre-Cambrian rocks at greater depth are less dense. The hypothesis of crustal warping appears to explain logically both the negative anomalies of the Valley-and-Ridge province and the positive values on the Piedmont, since there is geologic evidence of secular uplift and erosion in the latter

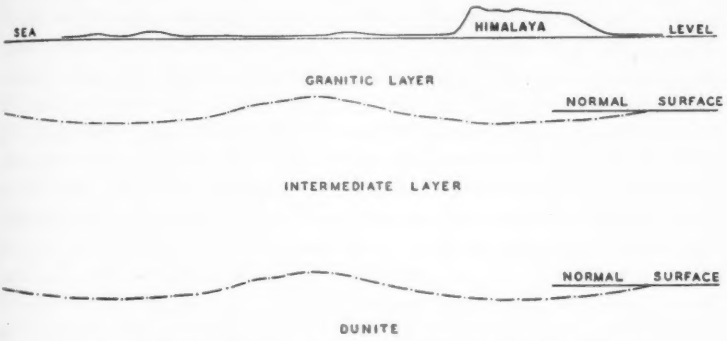


FIGURE 11.—*Hypothetical section in India showing Glennie's concept of crustal warping as an explanation of gravity anomalies*

Vertical scale greatly exaggerated.

province while the geosynclinal floor was sinking. Presumably the structural relief was greatly increased as a result of the Appalachian orogeny.[9]

CONSIDERATIONS OF ISOSTATIC BALANCE

If the segment of the crust represented in Figure 11 were divided into numerous columns equal in cross-sectional area, with the vertical dimension of each unit extending from the outer surface to some level in the dunite, these columns would not be equal in mass. The heaviest block in the plane of the section would lie along the axial plane of upwarp; the lightest blocks along the axial planes of downwarp. Hence Glennie considered his hypothesis of crustal warping incompatible with the doctrine of isostasy.

It would be possible by certain changes in Glennie's section to make all vertical columns of equal cross-sectional area equal also in mass. This could be accomplished by raising the base of the intermediate layer in proportion to the depression of its upper surface, thus compensating defect of mass in the upper part of any block by a corresponding excess at greater

[9] L. L. Nettleton in a recent paper (1941) concludes that the outstanding anomalies in the Appalachians of Pennsylvania have their source "in mass distributions rather deep in the crust, which must have their origin in the tectonic forces which caused the Appalachian uplift. . . ."

depth. Differential effects on surface gravity would be maintained, since the upper interface of the intermediate layer is closer to the stations than is the top of the dunite layer. However, such an assumed selective thinning of the intermediate layer would be highly artificial and difficult to defend on any logical basis. It is more reasonable, on geologic grounds, to assume regional instead of local isostatic balance in the crust (Daly, 1940). Columns that have a lateral extent equal to half the wave length of warping would be approximately equal in mass. The Glennie hypothesis does not appear to be closely reconcilable with the Airy concept of compensation by variable thickness of a light crust. However, it may be combined with a compromise between the Airy and Pratt concepts, if the assumed balance is between columns of large dimensions. The columns containing a segment of the Himalaya Mountains (Fig. 11) may be in large measure compensated according to the Airy scheme. Topography over the assumed downwarp in southern India has moderate relief and does not require for its support a "root" of sial as large as that indicated in the diagram. The problem of compensation in that column is difficult unless we assume a special distribution of densities at depth or adopt the concept of a strong crust with broadly regional compensation.

The crustal warping assumed for southern New England and eastern New York is on a smaller scale than that postulated for India, and the problem of isostatic balance is correspondingly less. Assuming that the width of columns in balance is half the wave length of the regional warping, the radius of a unit column may be no more than 50 or 60 kilometers. However, since the former surface irregularities of the region, once probably large, have been reduced to slight relief, it does not appear that compensation according to the Airy concept can play any important part between columns of moderate extent within the region. These considerations emphasize an important question that confronts the student of mountain history: If a large mountain unit results from thickening of a light crust, with a large downward bulge to compensate the high topography, in what measure is the bulge "floated up" and eliminated while the relief is being reduced by erosion? It is reasonable to suppose that a large remnant of such a bulge will remain after the mountain mass is peneplaned; stresses due to buoyancy of the thickened sial will steadily decrease until they are smaller than the strength of the crust. Such a remnant, under an old mountain belt like the Appalachians, may be registered in negative anomalies that have no necessary relation to the isostatic condition of the belt as a whole. A group of anomalies that has this origin should remain unresponsive to corrections based on the usual assumptions of isostatic compensation or on specific gravities of the surficial rocks. These characteris-

tics are conspicuous in the groups of anomalies analyzed in the present study.

The spacing of "low" and "high" axes in the New England-Hudson Valley region, with a wave length of 100 to 150 miles, suggests a more or less rhythmic character in the influence responsible for the alternating groups of anomalies. Such a character is logically associated with a series of broad, gentle folds affecting the crust. The maximum size of these assumed folds or warps that can be maintained by the strength of the crust through geologic eras is an interesting subject. Presumably the shorter the wave length, the greater the amplitude of a downwarp that could endure after all compensating surface irregularities have disappeared through erosion. If other groups of anomalies are found that appear generally similar to those considered here, a comparative study taking account of wave lengths, gravimetric amplitudes, and other elements should yield interesting results.

CONCLUSION

The anomalies of the Hudson Valley-southern New England region contain elements of two general kinds: (1) that which depicts a regional tendency, not directly related to the visible geology; (2) that which clearly reflects visible rock masses characterized by abnormal densities. The two anomaly elements are combined in varying proportions, and in parts of the region that have complex bedrock it is not possible to separate them quantitatively. For a few bedrock units, notably the Triassic area in Connecticut and Massachusetts, anomalies can be partly analyzed and a correction can be applied for local influences, giving the regional tendency clearer expression. Outstanding groups of Bouguer anomalies with like sign are essentially unchanged by isostatic reduction, suggesting either actual departures from isostatic compensation or strong influence of concealed differences in density. The two principal groups of negative anomalies are in the Hudson Valley belt and the Rhode Island basin, both of which were areas of subsidence, sedimentation, and severe orogeny in the Paleozoic era. The outstanding group of positive anomalies lies between the two negative groups in an area characterized by relative uplift. Glennie's hypothesis of crustal warping, based on a study of India, offers a plausible explanation of the present problem. It is suggested that downward bulges at the base of the sial shell, formed in Rhode Island and eastern New York during the Paleozoic history of subsidence and mountain making, were not entirely eliminated during the later long history of uplift and erosion. This hypothesis is consistent with the concept of regional as opposed to local isostatic compensation.

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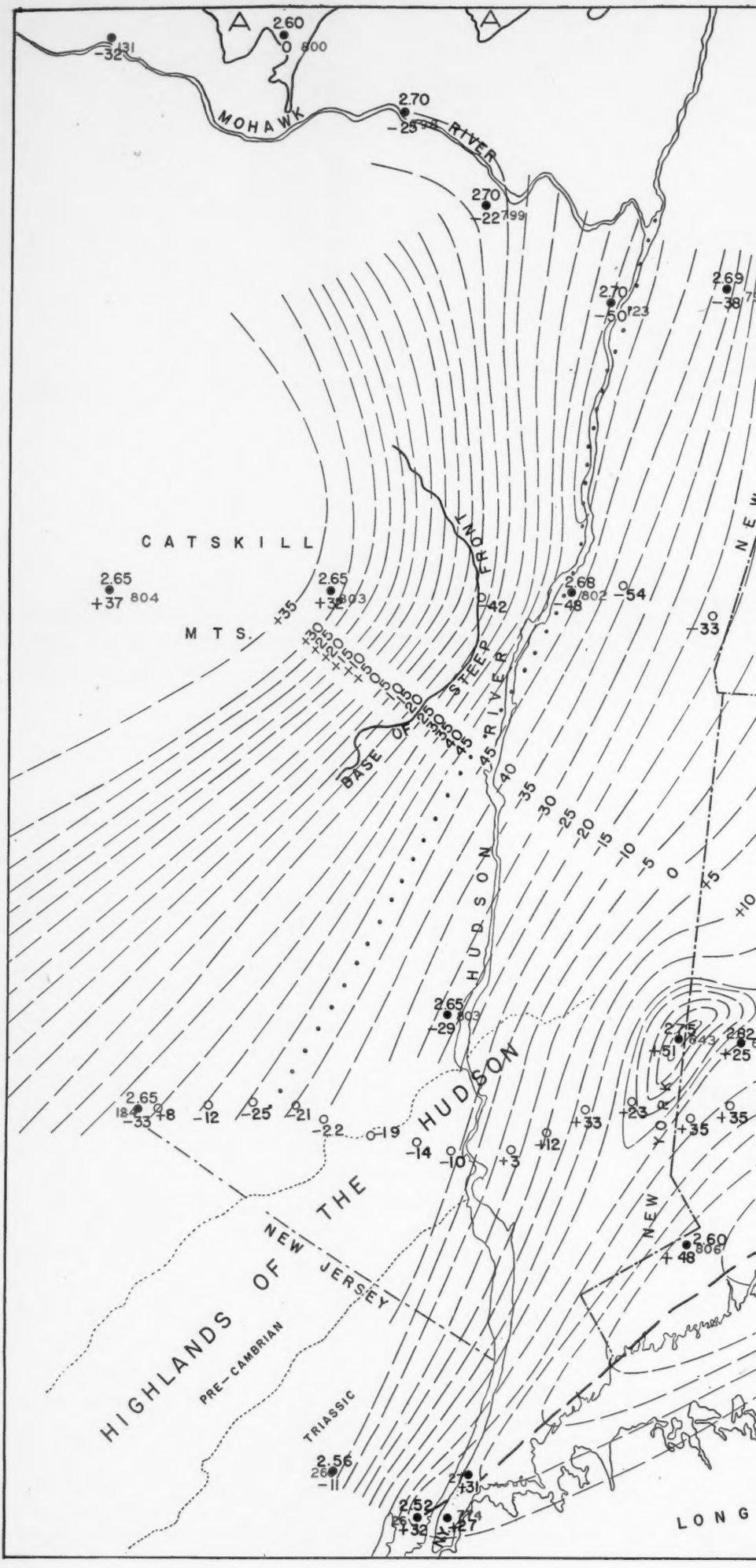
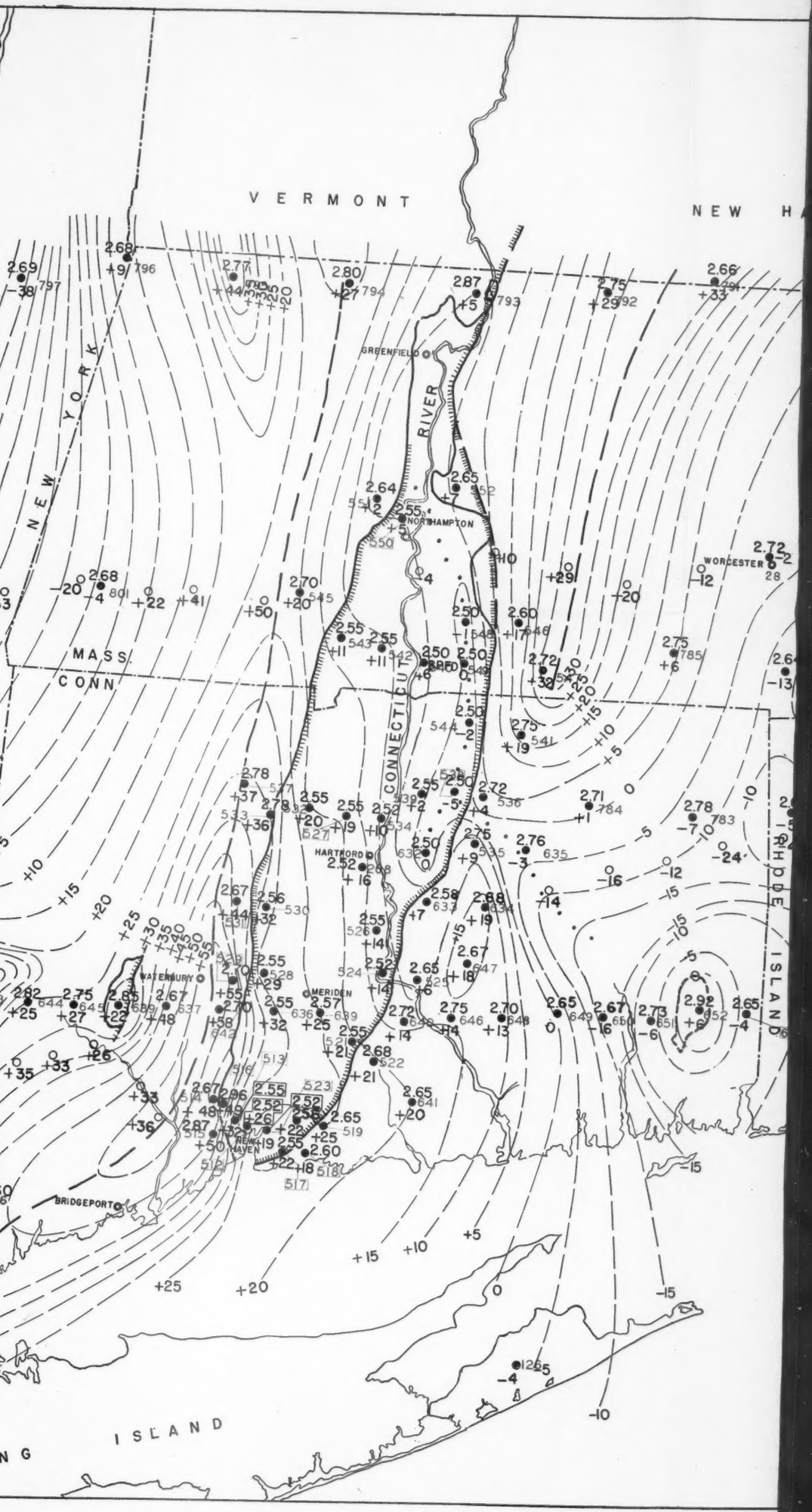
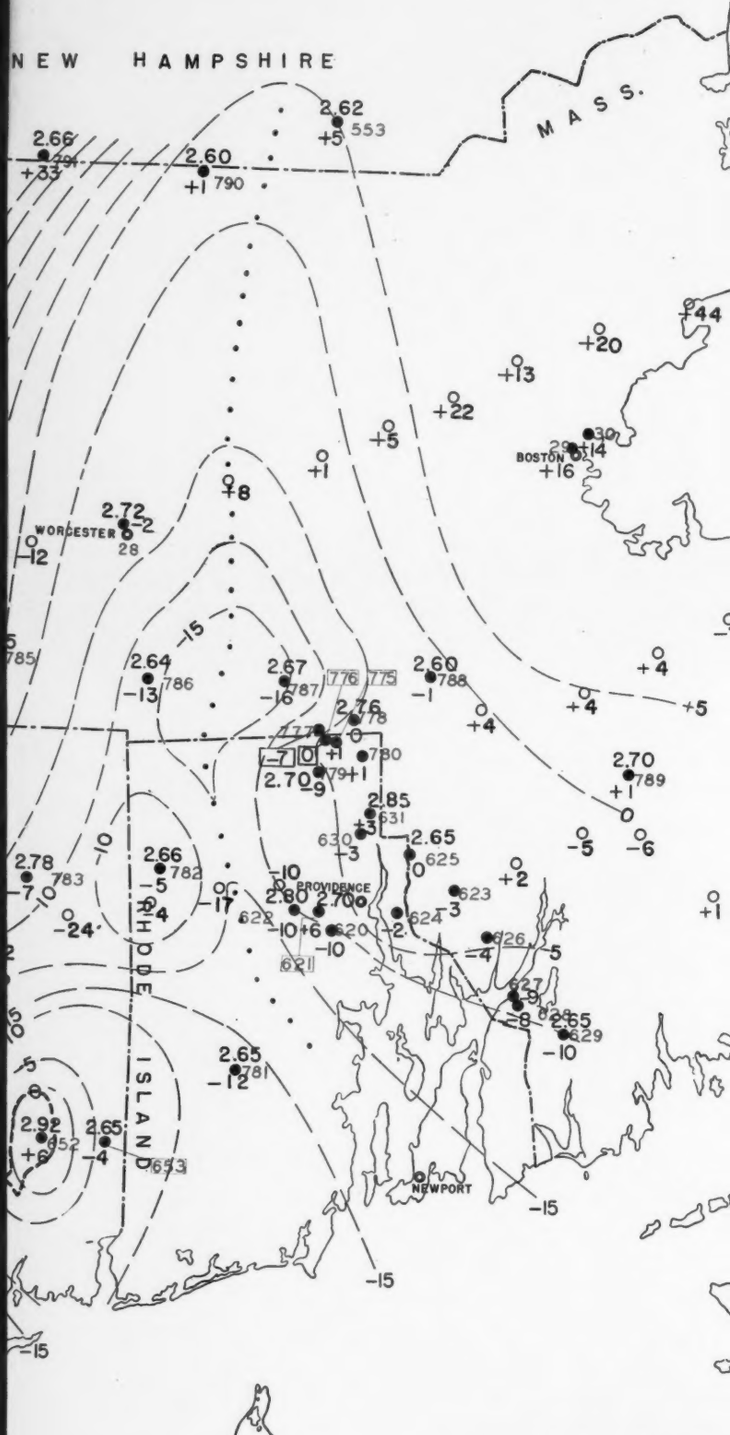
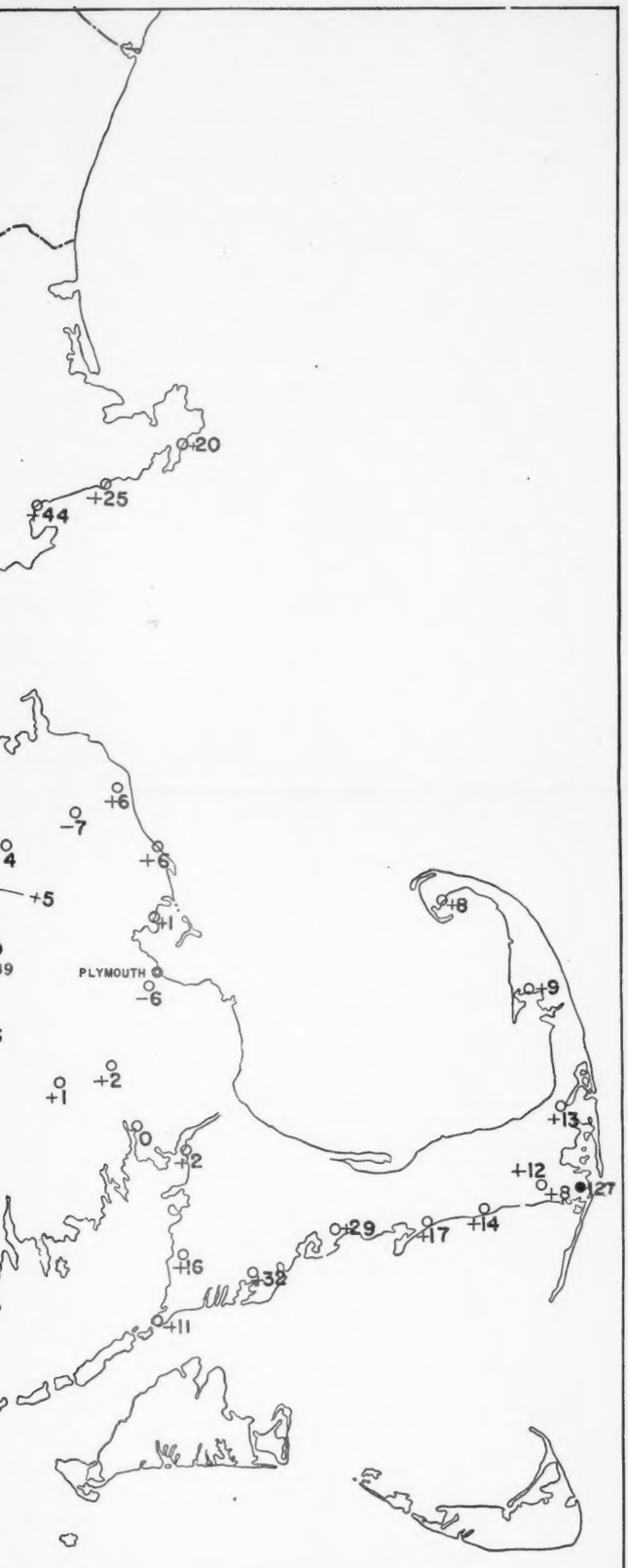

MOHAWK RIVER
CATSKILL MTS.
BASE OF STEEP
HUDSON RIVER
HUDSON
NEW JERSEY
HIGHLANDS OF THE
PRE-CAMBRIAN
TRIASSIC
LONG



ISANOMALY MAP, BASED ON FREE-AIR ANOMALIES


NEW HAMPSHIRE
MASS.
2.62
+5
553
2.66
+33
2.60
+1
790
+44
+20
+13
+22
+5
+1
+8
BOSTON
+16
2.72
WORCESTER
28
-12
-15
2.64
-13
786
2.67
-16
787
776
775
2.60
-1
788
+4
+5
2.76
778
777
-7
780
2.70
789
+1
2.85
631
630
+3
2.65
625
2.78
783
-7
2.66
782
-5
-24
-17
-10
2.80
2.70
PROVIDENCE
623
624
626
-4
627
628
629
-10
2.65
781
-12
2.92
652
+6
2.65
-4
653
RHODE ISLAND
NEWPORT
-15
-15
-15
ISANOMALY MAP, BASED O
FIGURES SHOW ANOMALIES
AVERAGE SPECIFIC GRAVI
● = PENDULUM STATIO
○ = GRAVIMETER STATIO
BUT ISANOMALIES NO
SPECIAL SYMBOLS EXPLA
0
25
MIL



ASED ON FREE AIR ANOMALIES.
OMALIES IN MILLIGALS, AND
GRAVITIES OF ROCKS NEAR STATIONS.
STATIONS (BASIS OF MAP)
STATIONS (THOSE OF 1940 ENTERED,
LIES NOT ADJUSTED TO THEM).
EXPLAINED ON PLATE I.

25 MILES 50

FIGURE 1. MADISON AND BRAZER FORMATIONS IN LOGAN CANYON
View NE from mouth of Spring Hollow. Unnamed Devonian (?) beds below.

FIGURE 2. BRAZER FORMATION NEAR DRY LAKE
View NNW toward Wellsville Mountain. (Mm) Madison; (Pwe) Wells

CARBONIFEROUS FORMATIONS IN LOGAN QUADRANGLE

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CARBONIFEROUS FORMATIONS OF THE UINTA AND NORTHERN WASATCH MOUNTAINS, UTAH

BY J. STEWART WILLIAMS

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ABSTRACT

The Carboniferous rocks have been studied from Wells Canyon, Idaho, southward to Big Cottonwood Canyon in the northern Wasatch Mountains and eastward to Brush Creek Canyon in the eastern Uinta Mountains. Lithologic descriptions and faunal lists are given for the successions in the Logan quadrangle and vicinity, and in Weber, Big Cottonwood, Duchesne River, and Brush Creek canyons. Eastward in the Uinta Mountains the Madison formation grades into drab sandstones that interfinger with red beds. A thick and excellently exposed section of the Brazer formation at Dry Lake, Logan quadrangle, is described. The Brazer is traced eastward nearly to the Colorado State line. The Morgan formation extends southward from the type locality and eastward along both sides of the Uinta Mountains. A 5000-foot section of the Wells formation in Wellsville Mountain, west of Cache Valley, is described. The lithology and age of the Park City formation are discussed.

The lower part of the Morgan formation in Big Cottonwood Canyon is probably Bendian in age, the upper part Des Moines. The Morgan and Weber formations in the Uinta Mountains are Des Moines and are believed nearly synchronous with the lower part of the Wellsville Mountain exposure of the Wells and with the type section in Wells Canyon, although part of each of the last-named sections may be somewhat older (Bendian). Some evidence supporting a Permian age for the whole of the Park City formation is presented. The paleogeography of the area through Carboniferous time is reviewed.

INTRODUCTION

The Carboniferous rocks have been studied from Wells Canyon, Idaho, southward to Big Cottonwood Canyon and eastward to Brush Creek Canyon in the eastern Uinta Mountains (Fig. 1). Lithologic descriptions and faunal lists are given for the most complete and best-exposed sections in the area. The thick section in Big Cottonwood Canyon is reported for the first time, and the Madison, Brazer, and Morgan formations are distinguished in the Uinta Mountains sections as far east as Brush Creek. Thick sections of the Brazer and Wells in the Logan quadrangle and vicinity have been used to relate the Wells to the Weber and Morgan, and the type Brazer to the Brazer sections in the northern Wasatch and Uinta Mountains. Some additional evidence bearing on the question of the validity of the Park City formation as a stratigraphic unit has been gathered.

Although the writer had accumulated considerable information for several years, most of the work was done in the summer of 1941. Complete sections were studied in the Logan quadrangle and vicinity, and in Weber, Big Cottonwood, Duchesne River, and Brush Creek Canyons. The type localities in Brazer and Wells canyons were also visited.

ACKNOWLEDGMENTS

This study was made possible by a grant-in-aid from the Penrose Bequest of The Geological Society of America; the opportunity thus provided is greatly appreciated. The writer is indebted to Mr. Alvin M. Hanson for his excellent help as field assistant. Dr. M. L. Thompson was kind enough to furnish the age value of fusulinids from Wellsville Mountain.

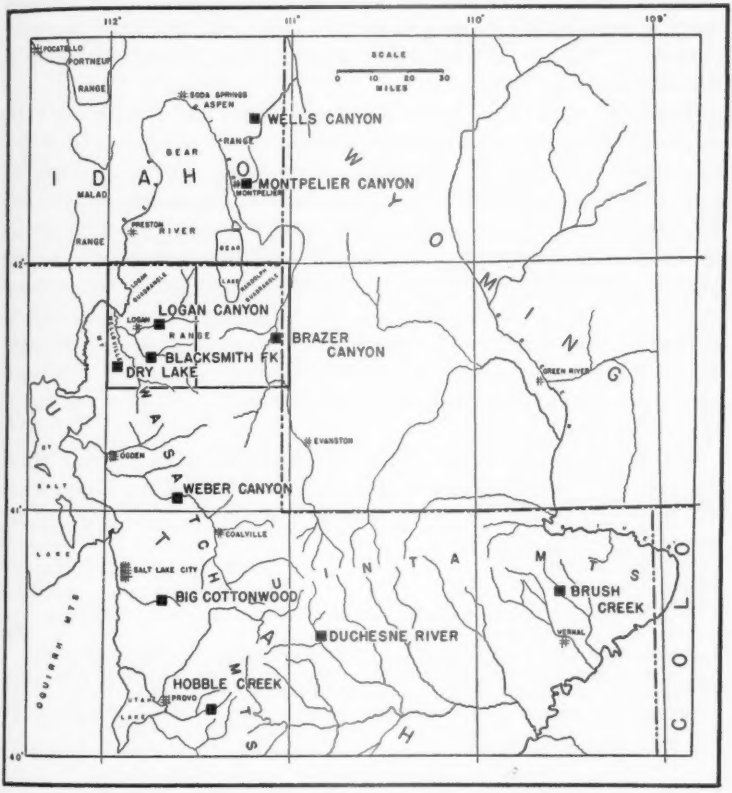

FIGURE 1.—*Index map showing locations of sections*

SECTIONS STUDIED

LOGAN QUADRANGLE AND VICINITY

The Madison, Brazer, and Wells formations are well developed and well exposed in the mountains around southern Cache Valley. They have

received some study in this area by previous workers. Bailey (1927) described briefly and mapped in a preliminary way the Paleozoic systems in part of the Bear River Range. Hammond (1930) and Hansen (1930), students of E. B. Branson and Bailey, measured and summarized the Madison formation, and another student, Gunnell (1932), made a similar study of the Brazer formation, under the same directorship. The writer published (1939a, p. 27) a detailed section of the Brazer formation in Blacksmith Fork Canyon.

The upper walls of the canyons of Logan River and Blacksmith Fork, in their lower parts, offer beautiful displays of the Madison and Brazer formations (Pl. 1, fig. 1). The outcrops are nearly horizontal where the axis of the Logan Peak syncline crosses these canyons, and the walls rise in a series of gray limestone cliffs separated by green slopes developed on the weaker members of the succession. The Wells formation, with a great thickness in this vicinity, heretofore unrecognized, forms the northeast slope of Wellsville Mountain. The latter, forming the western wall of southern Cache Valley, is a great homoclinal block striking northwest, in which the formations dip northeast. The Wells is excellently exposed across the crest and down the northeastern slope of the northern end of this mountain.

The following composite section has been taken from sections of the Madison in Logan Canyon, the Brazer in Blacksmith Fork Canyon, and the Wells at the northern end of Wellsville Mountain (secs. 10, 11, T. 11 N., R. 2 W., S. L. B. & M.).

WELLS FORMATION — Feet

8. Medium-gray to drab, medium- to thin-bedded calcareous sandstone weathering buff, olive buff, or brown. Fossils rare..... 2700

7. Medium-gray, light-gray and pale mouse-gray calcareous sandstone and sandy limestone, thin to thick bedded and weathering olive buff to buff brown, alternating with and grading into massive beds of dark neutral-gray limestone, weathering light gray. Sandy beds constitute about 80% of the succession, limestone beds the remainder. *Triticites* abundant from 1000 to 2000 feet above the base. In the basal 170 feet, intercalated with sandy limestones, are some beds of shaly limestone, weathering olive buff, with abundant cryptostomatous bryozoans, silicified, and weathering in relief. Fusulinids present with bryozoans..... 2300

Triticites sp.
Zaphrentid corals
Species A (large)
Species B (small)
Lophophyllum sp.
Syringopora sp.
Batostomella sp.
B. sp.
B. sp.
Chainodictyon sp.
Archimedes sp.
D. cf. D. portlockianus (Norwood & Pratten)
Linoproductus cf. L. platyumbonus Dunbar & Condra
L. prattenianus (Norwood & Pratten)
Dielasma bovidens (Morton)
Squamularia perplexa (McChesney)
Spirifer occidentalis Girty
Neospirifer dunbari King

	Feet
WELLS FORMATION—Continued	
Lyropora sp.	
Polypora sp.	
Pinnatopora sp.	
Ptilopora sp.	
Orthotichia schuchertensis Girty	
Derbya crassa (Meek & Hayden)	
Chonetes granulifera Owen	
Juresania	
Krotovia sp.	
Dictyoclostus cf. D. americanus Dunbar & Condra	
Ambocoelia planoconvexa (Shumard)	
Punctospirifer kentuckyensis (Shumard)	
Hustedia cf. H. multicostatus	
Chothyridina orbicularis (McChesney)	
Composita subtilita (Hall)	
Euomphalus sp.	
Total	5000
BRAZER FORMATION	
6. Mostly thick-bedded compact gray limestones with fossils, of which corals are in the great majority. Upper 160 feet light- or medium-gray crystalline dolomite.	830
5. Light-gray sandstones and quartzitic sandstones, thin to medium bedded, weathering buff and buff brown; and medium-gray and dark-gray sandstones and calcareous sandstones, thin to medium bedded; alternating with fuscous-black and medium-gray limestones, thin bedded to massive. About one-fourth of succession is limestone.	830
4. Phosphatic shale member: mostly black shale, with beds of fuscous-black compact limestone and brownish-gray sandstone.	75
Total	1735
MADISON FORMATION	
3. Dark-gray and medium-gray compact limestone, generally thin bedded, with abundant fossils. Black chert throughout this unit in stringers a fraction of an inch to several inches thick, but concentrated near the base and the top. Upper part dolomitic.	765
2. Thin-bedded shaly limestone and shale. Some beds of fuscous-black limestone.	60
1. Thin-bedded dark-gray to fuscous-black limestone. So-called "Contact Ledge."	20
Total	845

The Pisgah Hills, essentially a low ridge extending southeastward from Wellsville Mountain and separating small Mantua Valley from Cache Valley, is developed on the three Carboniferous formations. The most resistant member of the succession, the upper limestone member of the Brazer, forms the crest of the ridge. Along this ridge and northwestward upon the Wellsville Mountain (Pl. 1, fig. 2) the Brazer formation has changed noticeably from its composition in Blacksmith Fork. Here it is much thicker, 3100 feet, and has developed a thick member of shales and silty limestones above the thick-bedded limestones. The sandstone-limestone member is thin, and the basal phosphatic shale member is missing. Following is the section measured (except for the lowest member) across the hill (Pl. 1, fig. 2, foreground) at the north boundary of sec. 28, T. 10 N. R. 1 W. Logan quadrangle. This section is conveniently known as the Dry Lake section.

Feet

5. Mostly thin-bedded fuscous-black to gray silty limestone and calcareous siltstone weathering generally pale purple drab. Some fuscous-black platy shale weathering olive buff. Also some dark-gray thick-bedded compact to crystalline limestone. About 100 feet above the base of the unit is a thin member with abundant silicified fossils, most commonly *Chonetes loganensis*. These beds lie across the summit between Dry Lake and the head of Sardine Canyon..... 950

Schuchertella	*Spirifer brazerianus* Girty
Orthotetes kaskaskiensis Mc-Chesney	*Spirifer pellaensis* Weller
Chonetes loganensis Hall & Whitfield	*Composita subquadrata* Hall
Productus elegans Norwood & Pratten	*C. trinuclea* Hall
	Aviculopinna cf. A. nebraskensis Beede

4. Medium- to thick-bedded crystalline to compact dark-gray limestones with considerable smoky-gray and black chert. Lower 200 feet thinner bedded, with much smoke-gray chert showing Liesegang banding. *Canina* common throughout. Top of this unit forms the crest of hill..... 950

Canina sp.	*Chaetetes* sp.
Zaphrentis excentrica Meek	*Michelinia* sp.
Z. stansburyi Hall	*Dictyoclostus inflatus* (McChesney)
Lithostrotionella sp. A.	*Spirifer brazerianus* Girty
Lithostrotionella sp. B.	*Composita trinuclea* Hall

3. Thin- to medium-bedded grayish-olive argillaceous limestone weathering olive buff, alternating with thin units of medium- to thick-bedded compact to crystalline limestone. In some parts the silty limestone grades into fuscous-black shale weathering olive buff. High in this unit is a bed of limestone packed with shells of *Gigantella brazeriana*..... 470

Pentremites subconoideus Meek	*Gigantella brazeriana* (Girty)
Chonetes loganensis Hall and Whitfield	*Pugnoides uta* (Marcou)
Echinoconchus cf. E. alternatus (Norwood and Pratten)	*Spirifer rockymontanus* Marcou
Dictyoclostus inflatus (McChesney)	*Punctospirifer spinosa* Norwood and Pratten
Linoproductus ovatus (Hall)	*Composita sulcata* Weller
Productus elegans Norwood and Pratten (?)	*Myalina cf M. congeneris* Walcott
	Pleurotomaria pealeana Girty
	Paraparchites cf P. carbonarus (Hall)

2. Thick-bedded medium- to dark-gray crystalline to compact limestone. *Lithostrotion whitneyi* common..... 400

Lithostrotion whitneyi Meek	*Syringopora* sp.
Lithostrotionella sp. A.	*Gigantella brazeriana* Girty

1. Drab slightly calcareous sandstone that weathers to wood brown or cinnamon drab. Upward this unit grades through alternations of sandstone and limestone into the gray limestones above. "Worm" fossils common in the sandstone..... (est.) 300

Total 3100

WEBER CANYON

Dipping eastward at about 45°, the five Carboniferous formations of the northern Wasatch Mountains are well exhibited in the walls of Weber Canyon from Morgan to a point about 1½ miles below Croydon. To the geologists of the 40th Parallel Survey, this was the type section of their Wasatch limestone and Weber quartzite (King, 1876). Emmons

(Hague and Emmons, 1877, p. 384-388) described in detail the members of the succession, a description readily followed by anyone who examines the section. The estimates of thickness given by these men are uniformly too high, however, generally about double the measured thicknesses. Thirty years later in connection with studies of phosphate reserves, Weeks and Ferrier (1907, p. 454-455) described very briefly the Park City without giving thicknesses. After Hague and Emmons the major contribution was made by Blackwelder (1910a) who recognized the red beds below the Weber quartzite as a distinct formation, to which he applied F. B. Weeks' manuscript name, Morgan. He reported a nonmarine red-beds member in the Mississippian limestone and described unconformities at the base of the Morgan formation and beneath the Park City formation. The latter unconformity, according to Blackwelder, was profound and brought the phosphatic shale member of the Park City onto the Mississippian limestones 4 or 5 miles northwest of Weber Canyon (presumably on Cottonwood Creek). The same year he (1910b, p. 545-547) gave a detailed description of the Park City in connection with his studies of phosphates.

Recently A. J. Eardley has mapped a large rectangular area that includes Ogden in its northwest corner and Coalville in its southeast corner, and hence the whole of Weber Canyon as far as Coalville. Two of his students have studied the Weber formation in detail. Through the kindness of Eardley the results have been made available to the writer in advance of their publication.

Before discussing the findings of these previous workers in view of the writer's results, a summary of the Carboniferous section in Weber Canyon is presented.

Park City Formation	Feet
15. Light neutral-gray fine-grained sandstone, thin to medium bedded. Much chert. A distinct ledge former. Rex chert member of the Phosphoria formation.	170
14. Thin-bedded, nodularly bedded fine-grained sandstone	140
13. Upper phosphatic member; phosphatic shale member of the Phosphoria formation. Black shale, brown calcareous siltstone, and some dark-gray and black oölite, grading upward into fine-grained gray sandstone	40
12. Light neutral-gray limestone, dark neutral-gray and quaker-drab calcareous siltstone, and light mouse-gray calcareous siltstone	575
11. Lower phosphatic member. Fuscous-black shale	8
10. Light mouse-gray, neutral-gray, and dark neutral-gray limestone	65
9. Neutral-gray calcareous sandstone. Massive	65
8. Buff, slightly calcareous sandstone, sand poorly sorted, most of the grains well rounded. Some beds conglomeratic with pebbles of dark, tough quartz. At the base a grit composed of white chert fragments. Higher up this member is a massive sandstone intraformational breccia or conglomerate, varying in color from reddish to deep livid brown. Some limestone beds midway of the member	230
Total	1300

WEBER FORMATION

	Feet
7. Principally light-gray (smoke-gray to buff) sandstone, mostly medium grained, some quartzitic, some slightly calcareous, medium bedded to massive. About midway two comparatively thin members of gray and black chert (90%) in smoke-gray finely crystalline dolomite. In lower third thin beds of dark-gray limestone with fossils	3300

Zaphrentid corals
Aulopora sp. A.
Derbya **sp.**
Krotovia
Juresania nebraskensis (Owen)?
Dictyoclostus cf D. americanus Dunbar and Condra

Total (after Eardley) 3300

MORGAN FORMATION

6. Thin to thick-bedded, fine- to medium-grained earthy sandstones (mottled buff through flesh color to near kaiser brown). At the base some intraformational conglomerate and some paper shale. In the lower part the more earthy beds, thinly bedded and less resistant, bring into relief the more sandy beds, thickly bedded and more resistant. Upper beds are more uniform in composition and more evenly bedded, tending toward massiveness	1060
Spirifer occidentalis Girty	
Total	1060

BRAZER FORMATION

5. Neutral-gray and light neutral-gray compact limestones, thick to thin bedded, with shaly partings. The shaly beds weather olive buff	210

Lingula
Orbiculoidea
Dictyoclostus inflatus (McChesney)
Gigantella brazeriana (Girty) (small var.)?
Linoproductus ovatus (Hall)
Spirifer brazerianus Girty
S. pellaensis Weller
S. rockymontanus Marcou
Euomphalus sp. A.

4. Dark-gray limestones, thick bedded, with black shale partings	275

Zaphrentid corals
Aulopora sp. A.
Chaetetes sp.
Orbiculoidea
Orthotetes kaskaskiensis McChesney
Chonetes loganensis Hall and Whitfield
Echinoconchus sp.
Dictyoclostus inflatus (McChesney)
Productus elegans Norwood and Pratten
Dielasma
Spirifer brazerianus Girty
S. cf. S. pellaensis Weller
Eumetria
Composita trinuclea (Hall)

3. Some light neutral-gray, some neutral-gray, but mostly dark neutral-gray limestone, from compact and pure to sandy (95%); toward the top, two beds of thin-bedded buff sandstone, the upper causing a distinct gully that shows on both sides of the canyon (5%)	450
2. Light-gray to buff sandstones, some calcareous, some quartzitic, weathering without change of color or to some shade of brown, and in thin to thick beds; alternating with thin to massive beds of medium-gray, dark-gray or quaker-drab dolomite, limestone, and calcareous sandstone	820
Total	1750

MADISON FORMATION

1. Medium- to dark-gray limestone, compact to crystalline, and thin bedded. Typical Madison lithology	755
Total	755

It is readily apparent that the Mississippian rocks in Weber Canyon are very similar to those in the Logan and Blacksmith Fork canyons, except that the basal phosphatic shale member of the Brazer is absent in the north wall of the canyon. Neither includes a terrestrial member. However, between the Madison formation and the Jefferson dolomite, from Weber Canyon northward, are rocks that were deposited in part in the littoral zone. In Logan Canyon (Pl. 1, fig. 1) they are 850 feet thick—smoke-gray, thin-bedded sandstones and tawny olive to buff thin-bedded calcareous sandstones, with much intraformational breccia, and with the casts of halite crystals, small ripple marks, and rain-drop impressions common.

In the Weber Canyon area Eardley has mapped these beds as the Three Forks formation. To the writer's knowledge no Three Forks fossils have ever been found in them. But regardless of their age and correct designation,[1] is seems certain that they constitute that which was mistaken by Blackwelder for "a non-marine member in the Mississippian limestone."

In interpreting the base of the Park City formation, Blackwelder was misled by faulty identification of the various members of the sequence. The Weber does not disappear northward by thinning but by passing beneath the cover of Wasatch formation between Weber Canyon and Cottonwood Creek. It is present on the South Fork of Ogden River, and in the Logan quadrangle its equivalent, the Wells, is at least 5000 feet thick. Blackwelder must have mistaken the basal phosphatic shale member of the Brazer, lying just above the Madison, for the similar member of the Park City. Farther northwest, he must have mistaken the sandstones of the middle member of the Brazer for those of the Park City.

BIG COTTONWOOD CANYON

The upper part of Big Cottonwood Canyon, where these studies were made, is included in the Fort Douglas quadrangle, Utah, T. 2 S., Rs. 2 and 3 E. The general strike of the Carboniferous formations across the area is northwest, with a northeasterly dip (Pl. 2). The base of the section—that is, the contact of the Madison limestone on the underlying Cambrian rocks—crosses the ridge between Mill D South Fork and Days Fork (Reade and Benson Ridge) near the common corner of sections 29, 30, 31, and 32. The Madison and most of the Brazer were measured on Reade and Benson Ridge. The section was continued on

[1] The rocks mapped as Three Forks formation by Richardson in the Randolph quadrangle (1941, p. 19-20) are apparently the fuscous-black limestone and shale members (1 and 2 Logan quadrangle) at the base of the Madison formation. The fossils he has listed are common in these beds in the Logan quadrangle, but they are not Three Forks fossils.

the ridge east of Butlers Fork (misprinted Buttes Fork on the quadrangle map), from the highway to the crest of the peak in the NE ¼ sec. 7. This traverse crossed the uppermost part of the Brazer, as well as the Morgan and Weber formations. To check the Weber, and include the Park City, the study was completed on the ridge between Beartrap Fork and Mule Hollow, mostly in section 16. The ridge paralleling this one on the east is Section Ridge, the type section for the Park City, Woodside, Thaynes, and Ankareh formations, distinguished by Boutwell in 1907.

In the summer of 1912 Hintze (1913) studied the structure and stratigraphy of an area centering at Alta. He mistook the fauna of the Madison limestone for a Devonian one and suggested that the limestones now recognized as the Madison be called the Benson limestone (Devonian). Loughlin (1913), working in the area the same summer, collected the same fauna which Girty correctly identified as Madison. For those beds above the Benson, up to and including limestones with what is now recognized as a typical Brazer fauna, Hintze suggested the name Reade formation (Mississippian). In the same year (1913) that Hintze's report was published, Richardson's paper, defining the Brazer formation, appeared. Hintze and Richardson had each given a different name to essentially the same rocks, but subsequent usage has greatly favored the term Brazer. As already mentioned, the limestones near the top of Hintze's Reade formation contain the typical fauna of the Brazer as developed in southeastern Idaho and in the Logan quadrangle, Utah. Furthermore, they are essentially the same lithologically, and at least since Calkins (1933) wrote a brief guide to the geology along the ridge between upper Big and Little Cottonwood canyons the name Brazer has been applied to these rocks in Reade and Benson Ridge.

The writer could not validate the great unconformity distinguished by Hintze at the top of his Reade formation but measured, above this horizon an additional 1500 feet of sandstones and limestones, which he believes should be included in the Brazer formation, followed by 3300 feet of red beds, the Morgan formation, below the base of the Weber quartzite. As far as known this is the first recognition of this part of the section except as Boutwell (1912) reasoned that the underlying parts of the Weber, not exposed in the Park City district, must total at least an additional 1500 feet.

The Carboniferous section in Big Cottonwood Canyon is as follows:

Park City Formation	Feet
21. Not examined in detail except to search below the phosphatic member for fossils and to note the presence of the breccia bed (8 Weber Canyon) at the base. Thickness	760

View into Mill A Basin. (Mm) Madison; (Mb) Brazer; (Pm) Morgan; (Pw) Weber. Overthrust fault puts Brazer on Morgan and Weber.

NORTH SIDE BIG COTTONWOOD CANYON

FIGURE 1. MADISON CLIFFS
East side of North Fork.

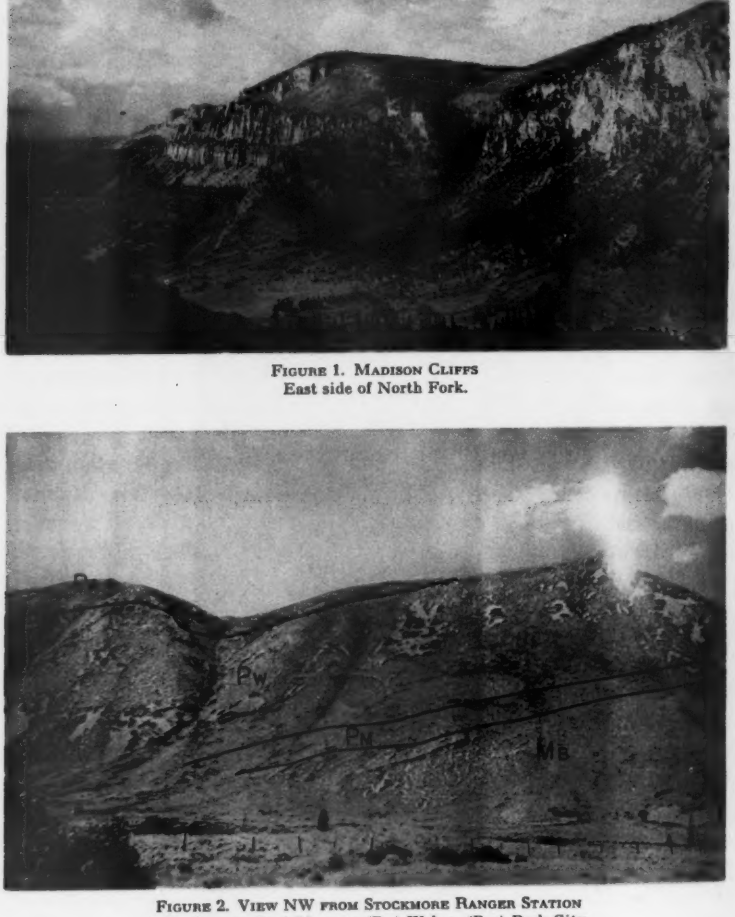

FIGURE 2. VIEW NW FROM STOCKMORE RANGER STATION
(Mb) Brazer; (Pm) Morgan; (Pw) Weber; (Prp) Park City

CARBONIFEROUS FORMATIONS, DUCHESNE RIVER

Weber Formation	Feet
20. Gray quartzite, medium-bedded	530
19. Two thinner members of thin-bedded neutral-gray, mouse-gray and smoke-gray cherty dolomites, chert black and brown and constituting from 30% to 60% of the rock; and some dolomitic sandstone: separated by a thicker member of light-gray quartzite. At the top the cherty dolomite grades into the unit of gray quartzite above	245
18. Mostly thick- to medium-bedded medium- to fine-grained light-gray sandstone, weathering pale pinkish buff. Some medium-gray thin-bedded calcareous sandstone weathering olive buff. Beds vary from no calcium carbonate in pinkish buff sandstone to medium-gray sandy limestones. Only a few beds quartzitic. Cross-bedding common. Toward the top of the unit, definitely more quartzitic	420
Total	1200

Morgan Formation	
17. Deep livid-brown sandstone and earthy sandstone; conglomerate of the same color with small fragments of gray limestone; also mahogany-red earthy sandstone and siltstones; also much shale of the same colors	335
16. Light-gray coarse-grained sandstone, grit, and small pebble conglomerate; quartzitic, cross bedded; massive	105
15. Brown sandy shale and thin-bedded sandstone; some micaceous; some with minute lenticular bedding. Essentially the same as the second unit above	895
14. Deep grayish-olive slightly calcareous sandstone as below	80
13. Cinnamon-drab thin-bedded sandstone and sandy shale. Some shale light drab and weathering to "tea leaves."	155
12. Grayish-olive sandstones and coquinas as below	230
11. Brown shales	170
10. Deep grayish-olive sandy limestone, weathering to a deep olive buff; alternating with beds of light-gray limestone, some massive. Part of the limestone is composed largely of shell fragments. Some reddish-brown intraformational conglomerate composed of small fragments of coquina and brown shale. This unit forms a conspicuous banded cliff. At the top of the unit the alternations are more frequent, and the succession grades into the weaker beds above *Branneroceras* sp.	410
9. Olive-buff and light-gray sandstones and brown sandy shales	925
Total	3300

Brazer Formation	
8. Thin-bedded limestones, most of them sandy or silty and with shale partings; much black shale in the upper 300 feet. Colors vary from white (some platy sandstone) through gray to dark drab. Shaly limestones weather olive buff	745
7. Thin- to medium-bedded gray calcareous sandstone (some limestone) alternating, in units of various thicknesses, with similarly bedded buff sandstone, often quartzitic. Fossils few. *Spirifer pellaensis* Weller *Euomphalus* sp. A	750
6. Mostly thick-bedded light-gray compact limestone and sandy limestone, with some chert. Some thin-bedded silty limestones that weather olive buff. Fossiliferous	225

Orthotetes kaskaskiensis (McChesney)
Spirifer pellaensis Weller
S. cf. S. pellaensis Weller
Punctospirifer
Composita
Cliothyridina hirsuta Hall

BRAZER FORMATION—Continued

	Feet
5. Quaker-drab silty limestone and calcareous siltstone with much black chert. *Caninia* zone.	40
Caninia	*Chonetes loganensis* Hall and Whitfield
Michelinia	*Dictyoclostus inflatus* (McChesney)
Pentremites godoni Defrance	*Productus elegans* Norwood and Pratten ?
Stenopora	*Spirifer brazerianus* Girty
Fenestella	*Punctospirifer*
Polypora	*Eumetria cf. E. verneuiliana* Hall
Thamniscus	*Composita*
Rhombopora	*Cliothyridina hirsuta* Hall
Orbiculoidea	*Sevillia*
4. Medium-gray medium-bedded limestone, some layers silty. Few fossils.	240
3. Dark olive-buff calcareous shale weathering amber brown. Also an unknown amount of black paper shale.	130
2. Buff and pinkish-buff fine-grained sandstones, some calcareous, some quartzitic, alternating, in comparatively thin units, with fine-grained gray dolomites and compact gray limestones.	835
Total	2960

MADISON FORMATION

1. Light-gray to dark-gray dolomite, thin to medium bedded, very cherty in parts. *Syringopora* common throughout.	1120
(From the unaltered masses)	*C. cf. C. loganensis* Hall and Whitfield
Schuchertella chemungensis (Conrad)	*Spirifer centronatus* Winchell
Schellweinella planumbona Weller	*Composita humilis* Girty
Chonetes logani Norwood and Pratten	*Euomphalus luxus* White
Total	1120

DUCHESNE RIVER CANYON

The canyon of the North Fork of the Duchesne River offers the most complete and accessible exposure of the Paleozoic sequence in the western Uinta Mountains. For this reason it has had considerable attention, and at least three previous workers have written about its stratigraphy. These earlier students, however, were so interested in the broad regional problems that none took time to examine in detail, and hence adequately describe, the rock units; for the same reason no one has recognized all the formations in the succession, or accurately correlated them with the section in the Wasatch Mountains.

About a mile upstream from the forks, outcrops of the Madison formation forming steep and picturesque cliffs emerge from the talus and slopewash at the base of the canyon walls and rise northward at a small and decreasing angle (Pl. 3, fig. 1). These are the most striking cliffs in this part of the canyon and cannot be overlooked. On the east side of the canyon, at the end of these cliffs, is a large slidden mass called locally "The Slope." Here the wall of the canyon, greatly steepened by Pleistocene glaciers, and supported only by relatively weak sandstones and sandstone breccias of the Brazer, has collapsed, and a large mass

of debris has moved downdip on top of the Madison ledges and turned out into the canyon.

The writer measured the section on the west side of the canyon, beginning at a point near the end of the Madison cliffs at the last place where the Pine Valley quartzite (Cambrian) showed through the talus. The traverse was carried southwestward up the slope to a *Syringopora*-bearing bed just below thick intraformational breccias. This bed was followed southward down the outcrop to the last small gully where it was visible, and from this point the traverse was continued to the top of the canyon wall. Subsequently study farther down the North Fork and in the mouth of the West Fork (Pl. 3, fig. 2) showed that this traverse had included all the Weber formation. This more southerly position, where the traverse was completed, is directly opposite "The Slope." That part of the canyon that has been described is in unsurveyed section 18, T. 1 N., R. 8 W., Uinta Base and Meridian.

The geologists of the 40th Parallel Survey did not traverse the North Fork of the Duchesne River in their initial visits to the Uinta Mountains (Emmons, 1907, p. 291). In fact their information on the stratigraphy of the range was not sufficient to enable correlation of the section there with that in the Wasatch Mountains, and the Uinta Mountain quartzite was equated to the Weber quartzite.

The passage of time seems to have focused attention on this correlation problem, with the idea that it could be solved by an examination of the Duchesne River section. Consequently, in 1903, Berkey and Boutwell visited the Canyon, followed by Emmons and Weeks in 1906. Berkey (1904), arguing merely on the basis of stratigraphic position, made a correct general correlation with the Wasatch section. Berkey interpreted the thick intraformational conglomerate at the top of the Madison as a basal conglomerate marking an erosion interval of considerable magnitude at the base of the "Weber quartzite." With this supposed erosional unconformity Berkey explained the discrepancy between the thicknesses of the Weber in Duchesne and Weber canyons, a difference not nearly so great as he supposed, since the Weber in Weber Canyon is only 3000 feet thick, not 6000 feet as then thought. This supposed unconformity he equated to that at the top of the Uinta Mountain group in the canyons of the Green River at the east end of the range. Emmons (1907) and Weeks (1907) agreed with Berkey on the general succession in Duchesne River Canyon and its correlation with the Wasatch section, but they could not find the basal conglomerate he described. Emmons maintained that if an unconformity did exist beneath the Weber quartzite it could not be equivalent to that beneath the "Red Wall" limestone in Green River Canyon since the "Red Wall" is equivalent, in part, to the "Wasatch

limestone," which lies below the Weber. Weeks gave a good general description of the Duchesne River section, clarified by a study of the stratigraphy throughout the whole Uinta Range. Forrester (1937) renamed the "Ogden quartzite" of all previous writers the Pine Valley quartzite and assigned to it a Cambrian age. He found no evidence of an unconformity between the Mississippian and Pennsylvanian rocks.

The section in the North Fork of the Duchesne River is as follows:

	Feet
PARK CITY FORMATION	
9. Detailed descriptions for section on Blind Stream, 4 miles east, published by writer in 1939 (b). Thickness.	230
WEBER FORMATION	
8. Mostly fine- to medium-grained sandstone in various shades of light gray and buff, mostly in thick beds. Some fine-grained tough medium-gray sandstone with fossil fragments; near the top some dolomitic sandstone...	850
Mesolobus mesolobus (Norwood and Pratten) *Dictyoclostus* *Spirifer rockymontanus* Marcou *Punctospirifer kentuckyensis* (Shumard)	
Total	850
MORGAN FORMATION	
7. Mostly thin-bedded kaiser-brown earthy sandstone and sandy shale, intercalated with thicker beds of purple-drab, pale-buff and white calcareous sandstone and quartzite. Some intraformational conglomerate. Cross-bedding common in the thick sandstone beds. Grades upward into the buff calcareous sandstones of the Weber formation.	215
Total	215
BRAZER FORMATION	
6. Upper half: medium- to thin-bedded light-gray calcareous sandstone and wood-brown sandy limestone with partings of sandy shale that weather olive buff; abundant fossils.	
Orthotetes cf. O. kaskaskiensis Weller *Girtyella* *Echinoconchus* *Dictyoclostus inflatus* (McChesney) *Gigantella brazeriana* (Girty) (small var.) ? *Spirifer pellaensis* Weller *S. cf. S. platynotus* Weller *Punctospirifer* *Composita subquadrata* Hall *C.* sp. *Connularia* sp.	
Lower half: light-gray and medium-gray compact limestones, about one-third white chert, in large masses.	130
5. Olive-buff medium-grained sandstone, drab fine-grained sandstone, and pale smoke-gray calcareous sandstone alternating irregularly in thin and thick beds with medium-gray and fuscous-black compact limestones. Southward down the outcrop at the river's edge where this unit forms conspicuous ledges, it contains much intraformational breccia as follows: massive beds of breccia composed of fragments of compact limestone and sandy limestone, light-gray and medium-gray, some olive-brown, and buff-brown medium-grained sandstone in a matrix of light-gray to buff medium-grained sandstone; limestone fragments predominate and are in all sizes up to several feet across; in much of the breccia there is little matrix; in places the breccia tends toward an edgewise conglomerate; the whole is rudely bedded, and there are some beds of compact limestone having the regional attitude.	370
Total	500

MADISON FORMATION — Feet

4. Light-drab medium- and fine-grained sandstone and calcareous sandstone, in beds that vary from massive to thin, intercalated with similar beds of light-drab sandy limestone and gray, compact, sometimes crystalline limestone. In the Upper 300 feet is much intraformational breccia with fragments of light-drab sandstone and gray compact limestone. Fossils common, particularly *Michelinia* and *Syringopora*. 915

Zaphrentid corals
Michelinia
Syringopora
Rhipidomella sp.
Schizophoria
Chonetes logani Norwood and Pratten
Chonetes loganensis Hall and Whitfield
C. sp. A.
Linoproductus cf. L. altonensis (Norwood and Pratten)
Spirifer centronatus Winchell
S. cf. S. forbesi Norwood and Pratten
Cliothyridina obmaxima McChesney

3. Thin-bedded limestone. Beds varying from laminae of near smoke-gray silty limestone to medium-thick beds of near quaker-drab crystalline and fossiliferous limestone. The silty or shaly beds predominate in lower part, the thicker beds in upper part. Like typical Madison lithology of sections farther west, except for the smoke-gray silty beds. Upper 30 feet thick-bedded crystalline limestone. 165

Schellweinella planumbona Weller
Rhynchotreta ?
Spirifer centronatus Winchell
Punctospirifer cf. P. transversa (McChesney)
Composita humilis Girty
Cliothyridina
Euomphalus

2. Grayer than mouse-gray very sandy limestone, sand grains large, well-rounded, and of translucent quartz, in part concentrated in laminae that weather in relief, showing cross-bedding. This sandy limestone, 4 feet thick, marks the base. Upward the sandy limestone grades into neutral-gray compact limestone with fossils. 27

Schuchertella cf. S. chemungensis (Conrad)
Chonetes sp. A.
Camarotoechia cf. C. metallica (White)
C. cf. C. gregarius Weller
Productella cf. P. pyxidata (Hall)
Avonia cf. A. concentrica (Hall)
Spirifer centronatus Winchell
Cliothyridina
Euomphalus

1. Smoke-gray and pale grayish-olive compact fine-grained sandstone, slightly calcareous, thinly sprinkled with large quartz grains. 8

Total 1115

BRUSH CREEK CANYON

Brush Creek Canyon (Marsh Peak quadrangle, Utah-Wyoming) affords the most accessible good exposures of the Carboniferous formations east of the North Fork of the Duchesne River and west of the Green River canyons. The gorge of Ashley Creek, a few miles to the east, has magnificent displays of these beds in its sheer walls, but it is extremely difficult to enter, and the walls are too steep to be conveniently worked.

A recently constructed road across the southern half of T. 1 S., R. 21 E. (S. L. B. & M.) connects the Manila-Vernal highway (U. S. 44) with the Taylor Mountain road. This new road crosses Brush Creek at the

base of the Carboniferous section—that is, at the contact of the Madison formation on the Pine Valley quartzite. About one-fourth mile downstream from this crossing Brush Creek disappears into a huge cavern developed in the intraformational breccias of the Madison formation. Just below the southern boundary of the township, where Davis Hollow joins the canyon, the outcrop of the Morgan red beds crosses the stream. From this point on through the massive sandstones of the Weber formation, Brush Creek Canyon is a narrow, rugged, almost impenetrable gorge. Therefore the Weber was measured along the road in Dry Fork, farther west. The writer had previously measured the Park City formation at the mouth of Ashley Creek gorge (Williams, 1939b, p. 98).

The reconnaissance report of the 40th Parallel geologists (Hague and Emmons, 1877, p. 290) sketches briefly some of the geological features of the broad terrace or shoulder of the Uinta Range in which the upper parts of the canyons of Ashley and Brush creeks are cut. Their map is one of four large-scale delineations of the geology of this area (Weeks, 1907; Schultz, 1919; Forrester, 1937). Since Powell's (1876) original description of the geology of the Green River canyons, Reeside (1925) has added notes on the section there exposed, but the writer knows of no detailed description of the Carboniferous succession in this area.

The traverse on which the formations were examined and measured runs southeastward from a point near the bridge over Brush Creek, along the road to the turnout at the beginning of the cave trail. The ridge followed by the trail down to the cave being essentially a dip slope, the traverse was broken at this point and resumed just west of the cave mouth at the base of the massive intraformational breccias in which most of the cave is excavated. From here it ran southwesterly to the top of the conspicuous gray cliff supported by the limestones of the Brazer. The beds beyond in this direction being covered, this horizon was followed down to the bottom of the canyon, and the measurement resumed up the east side where exposures are good in a large regional sheep trail. Measurements in this position carried the traverse to the base of the Morgan formation, which was followed to the canyon bottom once more opposite the mouth of Davis Hollow. The Morgan was studied on the east side of the canyon at this place.

Following is the section so obtained.

	Feet
PARK CITY FORMATION	
10. Measured at the mouth of Ashley Creek gorge in 1938 (Williams, 1939b, p. 98). Thickness.	245
WEBER FORMATION	
9. Thickness estimated from the exposure on Dry Fork.	1000

MORGAN FORMATION

	Feet
8. Mostly thin-bedded earthy sandstones between mars orange and burnt sienna. Upward these beds grade into salmon ocher sandstones, which are transitional to the buff sandstones of the Weber formation. At the top of the red beds, interbedded with buff sandstones, are thin beds (1 to 6 inches) of white coquina containing abundant tests of *Fusulinella* sp.	385
Fusulinella sp. *Lophophyllum* sp. *Chonetina flemingi* Norwood and Pratten *Marginifera muricatina* Dunbar and Condra *Spirifer opimus* Hall *S. rockymontanus* Marcou *Neospirifer cameratus* (Martin) *Composita elongata* Dunbar and Condra? *Composita*	
Total	385

BRAZER FORMATION

7. Gray compact limestones and quaker-drab compact limestones intercalated with medium-grained porous white sandstone and gray quartzitic sandstone.	200
6. Light mouse-gray, light neutral-gray and medium neutral-gray compact limestones in thick beds, with some light-gray calcareous siltstone and some light-drab fine-grained sandstone in lowest part. Near base a shaly limestone with fossils replaced by red chert.	500
Spirifer aff. S. brazerianus Girty *Composita trinuclea* (Hall) *C. subquadrata* (Hall)	
Total	700

MADISON FORMATION

5. Mostly gray clay shale, but some compact light-drab sandstone, some buff-brown calcareous sandstone, and some brown sandstone in shades between mars orange and burnt sienna. Mud cracks and ripple marks common.	275
4. Light-drab limestones, siltstones, and sandstones, generally thin bedded. Limestones compact. Throughout much of this interval these beds are jumbled into thick masses of intraformational breccia. Base of this unit near bottom of Brush Creek Cave.	235
3. Light-drab siltstones and fine-grained sandstones with some compact light-drab limestone, all distinctly thin bedded. A thin member of brown thin-bedded sandstone.	255
2. Above terra cotta, fawn brown, and purple-drab sandstones, thick to medium bedded, below olive-buff, fine-grained sandstone,	180
1. Mostly light-gray, light-drab, and light-buff sandstones and siltstones, thin to medium bedded, but with considerable light-drab compact limestone.	420
Total	1365

FORMATIONS

MADISON FORMATION

Lithology.—The Madison limestone formation is perhaps the most uniform in character, and hence the most easily recognized of the Carboniferous formations in the area studied. This is particularly true in the northern Wasatch Mountains and northward into Idaho, and, of course, less true in the Uinta Mountains where the limestones grade eastward into drab sandstones. Excepting the thin member of black shale near the base of the formation in the Randolph and Logan quadrangles, the forma-

tion in the northern Wasatch Mountains and the Bear River Range is composed wholly of limestone of generally uniform character.

The Madison often forms sheer, unscalable cliffs (Pl. 1, fig. 1; Pl. 3, fig. 1) but is nowhere massive. It is generally thin-bedded and sometimes medium-bedded, but never thick-bedded. It is almost always medium gray and varies in texture from compact to granular, but it is most often the latter, and is sometimes composed entirely of shells, shell fragments, and crinoid columnals. It is often cherty, and in Logan canyon the upper part is somewhat dolomitic, but this is exceptional, in areas removed from fault zones and intrusive bodies.

At the mouth of Providence Canyon in the Bear River Range (Logan quadrangle) where the Madison is truncated by the East Cache fault, the limestone is not only completely dolomitized, but over half its volume has been replaced by black chert in small subcubical masses distributed more or less uniformly throughout the rock. In Big Cottonwood Canyon the Cambrian as well as the Carboniferous limestones are in most places dolomitized and in some places marmarized by the close proximity of the Cottonwood intrusive bodies and the numerous dikes and sills that radiate from these bodies. Near the south boundary of sec. 18, T. 2 S., R. 3 E., where the Madsion was measured in the west side of Reade and Benson Ridge, it is changed completely to gray dolomite and is recognized aside from its stratigraphic position only by its thin bedding and the common occurrence of *Syringopora*, not completely destroyed by the process of dolomitization. At various places along the strike of the formation, however, masses of unaltered fossiliferous limestone prove that before it was changed by metamorphism the Madison formation in this area fitted the general description set forth above.

A marked change in lithology has occurred in the Madison on the Duchesne River, although it is still readily recognized not only by its stratigraphic position and fossils but by its thin bedding. The basal bed of the formation here is a gray, only slightly calcareous sandstone. Above this is gray sandy limestone grading upward into thin-bedded fossiliferous limestone, much like typical Madison farther west, except that the limestone beds are intercalated with beds of smoke-gray silty limestone. These sandstones and limestones constitute the basal 200 feet of the formation only. Most of the remaining 900 feet is drab sandstones and calcareous sandstones, with only minor beds of gray crystalline limestone. The upper 300 feet of the formation contains much intraformational breccia developed with fragments of light-drab sandstone and gray compact limestone. The formation here has become largely drab sandstone.

On Brush Creek the limestones have almost completely disappeared, and the formation consists essentially of light-drab sandstones and siltstones, with tongues of red beds and a thick member of intraformational breccia. This lateral gradation from gray limestones through drab sandstones to red beds, thick intraformational breccias developing with the sandstones and red beds, is clear evidence of a landmass in northwestern Colorado in Madison time that furnished detritus to the southeastern margins of the Madison sea.

Sloss and Hamblin (1942) have recently described in detail the stratigraphy of the Madison in Montana. It is interesting to note that a black shale member similar to that recorded by them for the base of the formation at a number of localities in Montana and North Dakota is present in the Randolph and Logan quadrangles, Utah. Since the Madison in the area treated by this paper is everywhere thin-bedded, it must be the equivalent of the Lodgepole limestone, and there is no correlative of the massive Mission Canyon limestone.

While the thickness of the Madison formation as recorded for this area (Fig. 2) varies, the variation is substantially less than that for the other formations listed, and it may be stated that the Madison maintains a thickness of about 1000 feet throughout the area.

Fauna.—That the fauna of the Madison limestone is Early Mississippian is well established, and the writer has no new information that in any way modifies this conclusion. It is thought, however, that a brief discussion of the fauna, based on the writer's collections from the formation in northern Utah, may be of value in distinguishing it from other Carboniferous limestones in adjacent areas, particularly the limestones of the Brazer.

Of the corals, *Syringopora*, including at least two species, seems the most abundant and widely distributed type. This genus occurs also in the Brazer and again in the Wells, but much less abundantly. The horn corals are smaller and less conspicuous in the Madison than in the Brazer. Of the Orthotetinae, *Schuchertella chemungensis* and *Schellweinella planumbona* are of frequent occurrence in the Madison, while in the Brazer *Orthotetes kaskaskiensis*, distinguishable from either by a median septum in its ventral valve, takes their place. *Chonetes* in the Madison limestones is generally smaller and more varied than those collected from the Brazer. They include two species, *C. logani* and *Chonetes* sp. A., that are distinctly more coarsely costate than any Brazer type. A *Camarotoechia* close to *C. metallica* is present in considerable numbers in nearly every collection of Madison fossils. *Spirifer centronatus* occurs far more frequently than any other *Spirifer* in this fauna.

The writer has always found the three common Madison gastropod species—*Euomphalus luxus, E. utahensis,* and *Straporallus ophirensis*—useful in distinguishing the formation. They are almost invariably present, at least one of them, in considerable numbers, on any outcrop. *E. luxus* is the most frequent.

BRAZER FORMATION

Type locality.—The Brazer formation was named by Richardson in 1913 for exposures in Brazer Canyon, Crawford Mountains, Utah. The geology of the canyon and mountains is delineated on the geologic map of the Randolph quadrangle, 1941. From this map it will be clear that neither the top nor the bottom of the formation is exposed in the Brazer Canyon section. Since this is the type locality of the formation, it seems worth while to publish the details of the section, which are as follows:

West Face of Mountain	Feet
Light-gray dolomitic limestone	10
Medium-gray dolomitic limestone in beds 6 to 12 inches thick	50
Medium-bedded, medium- to fine-grained buff calcareous sandstone	20
Medium-gray dolomitic limestone weathering somewhat lighter	25
Alternating beds of gray dolomitic limestone and buff, medium grained sandstone. Beds 2 to 8 feet thick, sandstone constituting three-fourths of the unit	20
Thick to thin-bedded light- to dark-gray dolomitic limestone, the alternating dark and light beds being 10 to 20 feet across	185
Alternating beds of sandstone and quartzitic sandstone, and light- to dark-gray dolomitic limestone. Beds 5 to 15 feet thick, the sandstone constituting about one-third of the unit	95
Dark- to light-gray dolomitic limestone	55
Gray dolomitic limestone. The uppermost few feet about half gray chert in thin stringers and beds	40
Fault	
Total	500

Very probably the limestones in this section are partly dolomitized because of their close proximity to fault zones, this alteration also accounting for the absence of fossils. The same is probably true of the whole belt of outcrop along the west face of the range, although Richardson (1941, p. 24) reports one small lot of fossils from this side, south of Rex Ranch. The section shows clearly, however, that it includes essentially alternating beds of sandstone and limestone. In Laketown Canyon, 15 miles northwest, a member of phosphatic shale lies at the base of the formation.

Lithology.—Westward in Blacksmith Fork and Logan Canyons the Brazer consists not only of an intercalated sandstone-limestone member and a phosphatic shale member, but includes a thick member of massive

limestones at the top (Pl. 1, fig. 1). Only 12 miles farther west, albeit across the major fault zones of Cache Valley, the formation has nearly doubled in thickness and is here mostly limestone, with considerable shale in the upper third, only thin sandstones near the base, and with no basal phosphatic shale (Pl. 1, fig. 2).

The exposure in Weber Canyon is closely similar to that in Blacksmith Fork Canyon, both in thickness and composition, except for the absence of the basal phosphatic shale member. The Cottonwood Canyon section, however, is again nearly 3000 feet thick, comprising, above an intercalated member, followed by a limestone and shale member, 750 feet of sandstone and calcareous sandstone. The uppermost part of the section here is silty limestone and shale. The thin sections on Duchesne River and Brush Creek both contain an intercalated limestone-sandstone member and a limestone member.

Despite the variety of lithologic types in the various exposures of the rocks carrying the Brazer fauna, the Brazer formation does possess some characteristic aspects. Every section contains at least one intercalated limestone-sandstone member, and every complete section includes a member of limestones, pure and generally thick-bedded. Finally, several of the successions contain considerable shale and shaly limestone, in addition to the basal phosphatic shale of the Blacksmith Fork Canyon and Laketown Canyon localities.

Certain specific rock types appear to be repeated in the various sections, as might well be expected. Units 5 Weber Canyon, 6 Big Cottonwood Canyon, 6 Duchesne River, and 6 Brush Creek, are very similar and appear at the top of the formation at two of the localities, Weber Canyon and Duchesne River. Unit 3 of the Dry Lake section is of somewhat the same type. The unit in which *Caninia* is abundant in the Dry Lake area (4) is quite like the *Caninia* bed of the Cottonwood Canyon area (5). It is doubtful, however, if any of these units are exactly synchronous, and their duplication simply means the recurrence of the particular conditions of deposition that brought them about, or the gradual shifting of the controlling environment from one area of the Brazer sea to another.

At every locality known to the writer the Brazer formation lies without angular discordance on the Madison limestone, and there is no evidence of erosion of the pre-Brazer surface, except in the variation in thickness of the Madison, which seems not too large to be accounted for by differences in rates of deposition in the Madison sea. The base of the Brazer is well marked by the phosphatic shale zone in the Bear River Range. In the Wasatch Mountains the first sandstone bed may be taken as manifesting the boundary. In the Uinta Mountains sections the drab

thin-bedded Madison sandstones contrast with the compact limestones, intercalated with buff sandstones, of the Brazer. The Morgan red beds, from Weber Canyon to Brush Creek, mark the top of the formation. In the Logan quadrangle the brown-weathering calcareous sandstones of the Wells, with fusulinids, are easily distinguished from the silty limestones of the Brazer, with *Productus*.

The Brazer varies greatly in thickness over comparatively short distances, as between Brazer Canyon, Blacksmith Fork, and Dry Lake; and Dry Lake, Weber, and Big Cottonwood, due apparently to erosion of the sediments of Brazer time prior to the deposition of the Weber and Wells formations.

Fauna and age.—Corals occur frequently. Species of *Lithostrotionella* seem most common throughout all the localities studied. *Lithostrotion whitneyi* is abundant in unit 2 of the Dry Lake section. *Syringopora* is much less frequent than in the Madison. *Pentremites* are readily found in unit 3 at Dry Lake and unit 5 at Cottonwood Canyon. *Orthotetes kaskaskiensis* and *Chonetes loganensis* are present at most outcrops. *Spirifer pellaensis* is by far the most abundant *Spirifer* in these sections, although *S. brazerianus* is often found.

Girty (Mansfield, 1927, p. 63-71) has discussed the age of the Brazer faunas at considerable length. To his general conclusion that the formation is the time equivalent of beds from the Spergen to the upper Chester the writer can add nothing. His suggestion of a possible zoning in the formation, with *Productus elegans, Camarophoria explanata,* and *Spirifer brazerianus* generally identifying the Chester horizon and *Gigantella brazeriana* or *Martinia lata* and *Lithostrotion* the pre-Chester zone, seems partly borne out in the Dry Lake section (Pl. 1, fig. 2), the only succession where fossils are general enough over a sufficient thickness to make an attempt at zoning at all feasible. Here *Lithostrotion* is abundant in the basal limestone, unit 2, and *Gigantella brazeriana* forms masses of limestone in unit 3, in which *Spirifer brazeriana* has not yet been found. The listing of *Productus elegans* from unit 3 would contradict this conception, but identification is not wholly satisfactory, since the specimens are few and not well preserved. Units 2 and 3 may represent part of the Meramec and St. Genevieve groups and units 4 and 5 may be correlative with the Chester group, but more must be known of the faunas before any such generalization can be proven.

It is interesting to note that *Archimedes* and *Lyropora,* two bryozoans characteristic of the Chester group, occur in the Wellsville Mountain sections in the basal Wells formation with fusulinids.

MORGAN FORMATION

Type locality.—Blackwelder (1910a) published the name Morgan, given by F. B. Weeks in an unpublished manuscript, for the red beds beneath the Weber quartzite in Weber Canyon. These red beds are conspicuous at the north side of Round Valley, a somewhat wider part of the Weber River Valley between Morgan and the Upper Narrows. Blackwelder recognized the terrestrial origin of these beds and, since they were unreported from other areas, suggested that they were probably of local extent only. Actually, they are over 3000 feet thick in Big Cottonwood Canyon and are developed throughout the Uinta Mountains.

A description of these beds at Round Valley is part of the description of the Weber Canyon section, already given. The thickness there given is a minimum only since the rocks are disturbed midway of the section, and some may have been faulted out. However, 1000 feet is probably close to the true thickness of the formation in this area.

Lithology.—The Morgan formation in Weber Canyon is almost entirely earthy sandstone, thinly bedded in the lower part but tending toward massiveness in the upper. Blackwelder described thin beds of limestone in places, which the writer did not see. A single marine fossil, however, was collected here.

In Big Cottonwood Canyon the Morgan red beds are 3300 feet thick (Pl. 2) and consist of brown shales grading through sandy shales into grayish-olive sandstones, intercalated with beds of grayish-olive sandy limestones grading into gray limestones. Upward in the succession the limestones disappear, and the upper half consists of shales and sandstones, with some pebble conglomerate. The upper 335 feet are "bright red" beds—that is, deep livid-brown and mahogany-red earthy sandstones and siltstones. Many of the limestone beds are coquinas of shell fragments; pelecypod shells are generally abundant in the limestones, but no other fossils were collected except the generally fragmented ammonoid shells from one horizon. These features, together with the presence of considerable intraformational breccia, and much cross-bedding in the sandstones, clearly indicate the shallow-water origin of the formation at this locality.

On Duchesne River the red beds below the Weber sandstone (Pl. 3, fig. 2) have essentially the same characteristics but contain no limestones. Here the Morgan is only 210 feet thick. On Brush Creek it is 385 feet, and here again thin beds of coquina are intercalated with the brown sandstones and shales, this time at the top of the formation. These coquinas are composed largely of the tests of *Fusulinella* sp.

The writer has observed the Morgan on Sheep Creek on the north side of the range in Daggett County; it is absent north of Weber Canyon.

These red beds lie without angular discordance on the Brazer formation. The writer has seen no evidence of pre-Morgan erosion other than the great variation in thickness of the Brazer. He was unable to verify the presence of an angular unconformity beneath the formation in Weber Canyon or the existence of a basal conglomerate at that locality, as described by Blackwelder. Upward in every section examined, the red beds grade into the buff sandstones of the Weber.

Fauna and age.—The fossils collected from the Morgan constitute strong evidence of an Early Pennsylvanian age for the formation. Those listed for the Weber, unit 8, Duchesne River, may be considered here since they came from sanstones at its base just above the red beds. Of those specifically identified in this list and that for the Morgan on Brush Creek, only *Chonetina flemingi* ranges above the top of the Des Moines series. (The identification of *Composita elongata* is questionable, and *Punctospirifer kentuckyensis* is a long-ranging form without significance.) The presence of *Fusulinella* confirms the Des Moines age of the beds. Both *Branneroceras* from the Cottonwood Canyon section and *Spirifer occidentalis* from Weber Canyon are Lower Pennsylvanian forms.

WEBER FORMATION

Lithology.—The Weber formation consists almost entirely of sandstone, with only minor amounts of limestone or dolomite. This sandstone is generally medium- to fine-grained and comparatively pure. The beds are nearly always thick, and sometimes massive, with cross-bedding common. Even at the type locality in Weber Canyon, only part of the sandstone is quartzitic; in the Uinta Mountains it is commonly friable.

Considerable beds of limestone are limited to the Wasatch Mountain sections. In Weber and Cottonwood canyons there are very cherty dolomites at about the middle of the formation, and at the former locality thin beds of dark-gray limestone are intercalated with the sandstones just above the Morgan red beds.

The Weber is not nearly so thick as once supposed, being only 3300 feet in Weber Canyon at its thickest exposure. From Cottonwood Canyon to Brush Creek it varies between limits of 1200 and 850 feet. Moreover, considering its clastic nature and its thick beds, often cross laminated, it seems certain that the formation was deposited in a comparatively short time.

At its base the Weber grades into the red beds of the Morgan, and

the latter clearly represents the initial phase of the deposition that later produced the Weber.

Fauna and age.—Only at two localities were fossils obtained from this formation. Those from just above the red beds in Duchesne River Canyon have been discussed with the Morgan fauna. Those from Weber Canyon are without specific stratigraphic value and bear no further comment. Since the age of the Morgan and at least the basal beds of the Weber in Duchesne River Canyon has been established as Early Pennsylvanian, and since, as already noted, the total thickness of the Weber probably does not represent any great length of geologic time, it may be reasoned that most if not all of the Weber formation is of Des Moines age, without excluding the possibility that the upper part of thicker sections may be younger than Des Moines.

WELLS FORMATION

Type locality.—A detailed description of the type section of this formation, in Wells Canyon, Crow Creek quadrangle, Idaho, appears in Mansfield's principal paper on parts of southeastern Idaho (1927, p. 72). The writer has studied the formation at this locality, but complete remeasurement of the section was unnecessary. It consists mostly of mouse-gray and buff calcareous sandstones, which grade into and are intercalated with beds of quartzite and limestone, in much lesser amounts. Lithologically, then, it does not differ essentially from the lower member of the Wells in Wellsville Mountain, Logan quadrangle (unit 7), or the Weber formation in Weber or Cottonwood canyons, except that the rocks are considerably less calcareous at the latter localities.

The beds at the top of the Wells as defined by Richards and Mansfield are particularly interesting since they would be part of the Park City of the Northern Wasatch Mountains (Boutwell, 1907) if we were to extend that name to this region. These beds have a better exposure on the South Fork of Deer Creek, on the west limb of the Webster syncline, about 1 mile to the northwest, than they have in Wells Canyon. Here, along the north wall of the canyon, thinner beds of buff calcareous sandstone, with some limestone, weathering to covered slopes, are followed by about 100 feet of buff calcareous sandstone which makes a conspicuous ledge on the canyon wall. Parts of this member are an edgewise intraformational breccia of calcareous sandstone fragments in a similar matrix. Above this conspicuous unit is an estimated 200 feet more of similar intraformational breccias and sandstone beneath the phosphatic shale member of the Phosphoria formation, although part of this thickness immediately beneath the shale, which is covered, may include the "underlying limestone" (Mansfield, 1927, p. 73). Southward in Montpelier Canyon, Mont-

pelier quadrangle, these beds are much better exposed, and the following were measured:

	Feet
FAULT	
PHOSPHATIC SHALE MEMBER, PHOSPHORIA FORMATION	
Grayer than mouse-gray fine-grained dolomitic sandstone or sandy dolomite. Heavy black chert beds at top. "Underlying limestone."	30
Intraformational breccia of buff sandstone	30
Intraformational (?) breccia of buff sandstone including fragments of red siltstone. Red fragments fewer toward the bottom. Thickness of unit varies greatly along strike	45
Intraformational breccia of buff sandstone	55
Total	160
TOP OF TALUS	

Much farther southward in Weber Canyon in unit 8 at the base of the Park City is a very similar massive breccia, part of a member of poorly sorted sandstone with a grit, composed of chert fragments, at its base. The breccia appears to be intraformational but the grit may well be interformational, the chert fragments having been transported from a land surface. The same or a very similar bed of sandstone breccia occurs at the same stratigraphic position in the Big Cottonwood Canyon section and also on the Duchesne River.

The presence of these beds of buff sandstone intraformational breccia (and interformational breccia in Weber Canyon), at the base of a stratigraphic unit, the Park City, that has a distinct topographic expression and, in its fossiliferous beds, a much younger fauna, argues strongly for the validity of Boutwell's formation. This will be discussed further in the next section, and it is sufficient to point out here that the writer would place that part of the type section of the Wells, including the intraformational breccia and the beds above it, in the Phosphoria (or Park City) formation. Thus restricted, the Wells formation at Wells Canyon is about 2000 feet thick.

Lithology.—The other Wells section with which this paper deals is that in Wellsville Mountain, southwest of Cache Valley. It has been described with the succession in the Logan quadrangle and vicinity, and the writer has already pointed out the similarity between the type Wells and the lower unit in the Wellsville section. The Wells is essentially calcareous sandstone, with some intergrading limestones and a little quartzite. In lithology it does not differ greatly from the Weber.

Fauna and age.—The only identifiable fossil collected by the writer from the Wells Canyon section was *Spirifer occidentalis* Girty from a bed about 400 feet above the base of the formation. While this single species is suggestive, its presence is hardly sufficient to date the formation, and

recourse must be taken to the Wellsville Mountain section, very similar lithologically, where a sizable sample of the fauna has been obtained.

The fauna listed for unit 7, Logan quadrangle succession, was gathered from the basal 400 feet of the formation, except only *Triticites* which appears about 1000 feet above the top of the Brazer. Of the fossils from these basal beds *Archimedes, Lyropora,* and *Ptilopora* are generally considered Mississippian; *Orthotichia schuchertensis* is Bendian; *Spirifer occidentalis* and *Cliothyridina orbicularis* are Des Moines; and the others definitely identified are all long-ranging forms of no stratigraphic significance except to indicate clearly the Pennsylvanian age of the formation; *Dictyoclostus americanus* and *Linoproductus platyumbonus* are limited to the Missouri series, but their identification in these rocks is only provisional. Fusulinids are present in these bottom beds and, when identified, should aid greatly in correlation, but the evidence now at hand seems clearly to indicate an Early Pennsylvanian age for the base of the Wells formation in Wellsville Mountain.

M. L. Thompson has examined the *Triticites* which are so abundant from 1000 to 2000 feet in this section. He reports that

> "they fit into the Pennsylvanian section of Kansas near the top of the Missouri series. Although they are not conspecific with any known Kansas forms, their evolutionary stage indicates this correlation. However, the Wells forms are very closely similar to an unnamed species from the Oquirrh formation."

He then explains that the Oquirrh species referred to comes from a horizon about two-thirds of the distance between that of the earliest fusulinids and that of the first Permian forms. It is surprising to find closely similar species of *Triticites* in these two formations which are lithologically similar, in the Oquirrh 10,000 feet up and in the Wells only 1000 feet up, the basal Wells having an Early Pennsylvanian fauna. This calls for a careful zoning of the fusulinids in the Wellsville Mountain section, but meanwhile it seems safe to conclude that the basal 2300 feet of this section includes rocks of Des Moines and Missouri age, perhaps even some of Bendian age.

Considerable risk would be involved in concluding that the 2000 feet of rocks in the Wells Canyon section cover this same time interval because they have essentially the same lithology as the basal 2300 feet in the Wellsville Mountain section. This opinion, however, is supported by Girty's faunal list for the formation in southeastern Idaho (Mansfield 1927, p. 73-75) and his own analysis of its age value.

PARK CITY FORMATION

Lithology.—At the type locality of this formation in Big Cottonwood Canyon three members may be distinguished—a lower member of sandstone with some limestone 400 feet thick, a middle member of phosphatic

shale and sandstone 35 feet thick, and an upper member of sandstone and cherty dolomite 340 feet thick (Hanson, 1942, p. 9-10). Eastward the lower member thins and disappears, so that the middle member rests directly on the Weber sandstone (Williams, 1939b); northward in Weber Canyon it is nearly 1000 feet thick, and it is present as far north as Wells Canyon, though much thinner, as has been pointed out. Each of the other members maintains a more uniform thickness in their Utah exposures and thickens somewhat in southeastern Idaho, where they constitute the Phosphoria formation. The lower member is lithologically similar to the upper members in that, in Weber Canyon, it includes an 8-foot bed of black phosphatic shale. The writer has already pointed out that the basal beds of this formation from Wells Canyon to Duchesne River Canyon consist of much intraformational breccia, evidence of a time of shallow water which, it is believed, initiated the deposition that continued until all the sediments of the Park City had been laid down. The phosphatic shales themselves may also represent shoaling of the seas and slow deposition, in keeping with Goldman's ideas (1922), but they are part of a lithologic unit which begins with the breccias and for which there is some evidence of faunal unity, as will be explained.

According to the statements of Mansfield and Girty (Mansfield, 1927, p. 78-79), use of the name Park City was not extended to Idaho because the lower member of the Park City formation, in that area, was found to contain Pennsylvanian fossils. The particular forms which were the basis for this decision are not listed. E. B. Branson (1916) likewise considered equivalent rocks in Wyoming as part Pennsylvanian and part Permian. The writer (1939b, p. 85-87), newly impressed with the continuity in lithologic character and thickness of the two upper members of Park City formation—that is, the Phosphoria formation—over a wide area including parts of Utah, Wyoming, and Idaho, and fortified by what was thought to be a bit of faunal evidence of his own, argued vigorously for the complete abandonment of the use of the term Park City.

A return visit to Holiday Park proved that the writer's first interpretation of the geology there was incorrect. The limestones from which the listed fauna came (1939b, p. 94) lie below the Weber quartzite, not above it, and hence have no relationship to the problem at all. During this study the lower member of the Park City was searched carefully for fossils at every locality with but little result. In Franson Canyon the calcareous sandstone at the top of the member yielded *Schizodus*, other pelecypods, and *Plagioglypta* similar to those that occur in the same lithologic facies of the upper member. This suggests that the whole formation was deposited without major interruption.

In other words, the evidence in this area, though hardly yet conclusive, now seems to indicate that the Park City formation is a stratigraphic unit of Permian age. This evidence may be summarized as follows: (1) the lithologic unity of the formation; (2) lithologic contrast with the Weber sufficient to produce definite topographic disparity at all points; (3) the presence of breccia beds, mostly intraformational, but perhaps in part interformational, providing the only known evidence of a prolonged break in deposition in the Morgan-Weber-Park City succession, at the base of the Park City; (4) the Early Pennsylvanian age of the Weber formation and Wells formation at Wells Canyon, compared to the Permian age of at least the middle and upper members of the Park City; (5) some evidence that the shallow-water pelecypod facies of the fauna of the upper member of the Park City also occurs in the lower member.

Additional evidence of the probable Permian age of the lower member of the Park City comes from adjacent areas. From the southern Wasatch Mountains (Hobble Creek Canyon) Baker and James S. Williams (1940) have reported an unmistakable Permian fauna from the 600 feet of limestone underlying the phosphatic shale member and probably the time equivalent of this member. Further, these fossils are hundreds of feet above the top of the Oquirrh formation from which *Pseudoschwagerina* has been obtained. From his work in central Wyoming C. C. Branson (1939, p. 1200) now considers the entire Park City formation (Phosphoria formation by his usage) a unit of Permian age.

While still discussing the lithology of the Park City the writer takes the opportunity to acknowledge the validity of Thomas' (1939) criticism of the terminology used in his 1939 paper with respect to the Mackentire tongue of the Woodside formation.

Fauna and age.—No data were collected during this study which contribute to the solution of the problem of the part of Permian time represented by the Park City and Phosphoria formations, generally considered Word in age.

AGE RELATIONSHIPS OF FORMATIONS

The sections of the Madison formation from one end of the area to the other are so uniform in lithologic character and fauna that there is little doubt but that the formation was deposited synchronously throughout the region. The drab sandstone facies, merging into red beds, that replaces the limestones east of Duchesne River, probably represents a somewhat shorter time interval for the same thickness but was deposited at the same time that pure limestones were accumulating farther from the shore of the Madison sea. With the Brazer, the question of what part of the thick Dry Lake section, and hence of Meramec-Chester time, is represented by

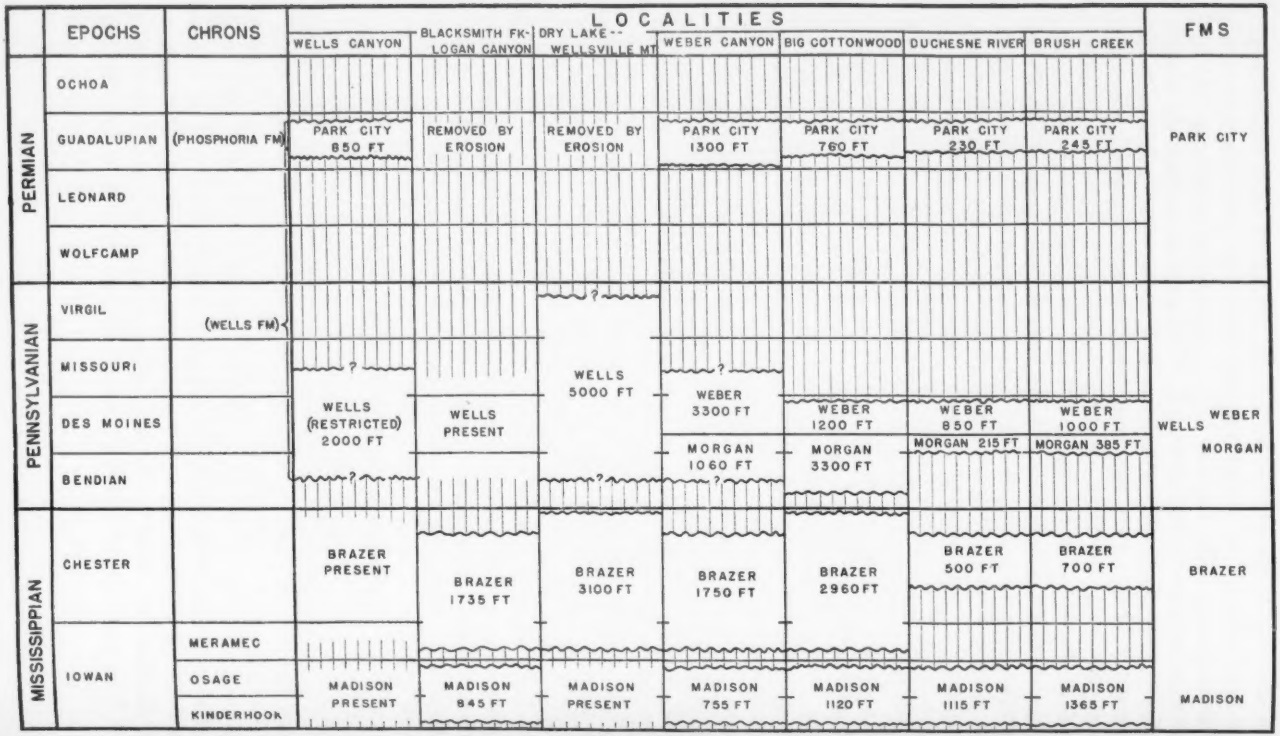

FIGURE 2.—*Age relationships of formations*

the thin sections of the Uinta Mountains is more difficult to answer. The faunas of unit 6, Duchesne River, and unit 6, Brush Creek, seem to correspond to that of units 4 and 5 of the Dry Lake section, however, and hence perhaps these units and the accompanying sandstones represent a part of Chester time. These faunas are characterized by *Orthotetes kaskaskiensis, Chonetes loganensis, Dictyoclostus inflatus, Spirifer brazerianus, S. pellaensis, Composita trinuclea,* and *C. subquadrata.* The *Caninia* zone, the lowest fossiliferous horizon in the Big Cottonwood Canyon section, likewise has a fauna of this same composition, as do units 4 and 5 of the Weber Canyon section. Pre-Chester time, then, may be represented by units 2 and 3, Weber Canyon, and units 2, 3, and 4, Cottonwood Canyon, and probably is not represented in the Uinta Mountains sections (Fig. 2).

The grayish-olive sandstones and intercalated gray limestones of units 10 to 14 of the Big Cottonwood Canyon section are probably earliest Pennsylvanian (Bendian). Units 15, 16, and 17 are similar, lithologically, to the red beds farther east and are probably Des Moines. Deposition of the Wellsville Mountain Pennsylvanian may have begun in Bendian time, though this is not yet certain. From the evidence known to the writer, the Wells Canyon section was contemporaneous with unit 7 of the Logan quadrangle succession, the lower unit of the Pennsylvanian recognized in Wellsville Mountain. All the Weber sandstone of the Uinta Mountains is probably Des Moines, but in the Weber Canyon area Missouri time may be represented in the upper part of this formation. On the site of Wellsville Mountain, deposition may have continued into Permian time as it did farther south in the geosyncline where the Oquirrh formation accumulated.

FIGURE 2.—*Age relationships of formations*

The phosphatic shale member of the Park City is apparently isochronal everywhere and may be used as a datum horizon. On this assumption deposition in Permian time began for this region first in the Weber Canyon area, soon spreading southeastward and northeastward to cover the Uinta Mountains area and southeastern Idaho.

PALEOGEOGRAPHY

Mississippian deposition in the area began with the precipitation, in places, of black shale along with the earliest limestones. Perhaps the sea spread into the region from the northeast, to become quickly a part of the shallow waters that covered most areas of the Cordilleran geosyncline in Kinderhook-Osage time. Toward the southeast, sandstones and red beds accumulated at the edge of the sea about a landmass that stood in northern Colorado and south-central Wyoming. Near the close of Madison time, either the waters withdrew first from this area, while the Mission

Canyon limestone was being deposited in Montana, or a very gentle upwarp at the very end of the Madison age permitted erosion of the Mission Canyon equivalent. Probably it was the former.

Deposition soon began again, the sediments being black phosphatic muds that accumulated in small basins scattered over a flat landscape about to sink beneath the encroaching sea. Meramec deposition probably was limited to the more persistent and hence earliest submerged parts of the trough, but by Chester time the sea had spread across the Uinta Mountains area and was there depositing some limestone. At the close of the Chester epoch general uplift drained all but the most persistently negative part of the trough, an area centering in the Oquirrh Mountains, but reaching over to the Big Cottonwood region and retaining a northward extension toward the site of Wellsville Mountain.

In this central part of the geosyncline, deposition seems to have been uninterrupted in the Oquirrh area and was not long suspended in the Big Cottonwood and Wellsville Mountain sections. In a short interval the sea spread northward to southeastern Idaho and eastward across the Uinta Mountains area. During the Des Moines epoch a thin unit of red beds, then a much greater thickness of cross-laminated sandstone, accumulated adjacent to the rising Colorado Mountains. Farther out in the trough, in the Wellsville Mountain and Weber Canyon areas, the sediments were at first partly limestones, but, as the mountains rose, the calcium carbonate deposited there was greatly diluted with current-brought sand. Toward the end of the Des Moines epoch restriction of the seas began, as an ever-increasing area rose with the Colorado Mountains, and by Missouri time only the Oquirrh-Wellsville Mountain area was still under the sea. Perhaps deposition continued into Virgil time in the last-named part of this persistent trough; certainly it did in the first-named part, where the sediments apparently accumulated without interruption from Mississippian into Permian time.

Though an arm of the sea existed from the Oquirrh area southward through the Wolfcamp and Leonard epochs, it did not spread northward again until Guadalupian time, when the Phosphoria sea developed over a wide area in parts of Utah, Idaho, Montana, and Wyoming. Deposition began early in the Weber Canyon area, where the sandstones and limestones of the lower member of the Park City are thickest, and by the time conditions became most favorable to the precipitation of phosphate the sea had spread eastward to the Colorado line and Central Wyoming (Thomas, 1934) where red beds interfinger with the shales and limestones. Accumulation of the sandstones, cherts, and limestones of the Rex member of the Park City or Phosphoria brought the Paleozoic record of the area to a close.

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RESTRICTED SUPPLEMENT

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RECORDED EXPERIMENTS IN THE PRODUCTION OF QUARTZ

BY

PAUL F. KERR AND ELIZABETH ARMSTRONG

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APRIL 1, 1943

EDITOR'S NOTE: Long-range scientific interest in the synthesis of sizable quartz crystals has risen to strategic importance under the present war emergency. As the first logical step toward any research in this field Doctors Kerr and Armstrong have completed an exhaustive search of the literature, classified and correlated all recorded experiments, and have submitted their compilation for publication in the hope that they might obviate the repetition of this effort by others and stimulate and assist additional experimentation.

The Geological Society is pleased to make this paper available to research workers in North America, but on account of its special character, distribution must be restricted. It is therefore issued in the format of the Bulletin but as a supplement thereto. Although it cannot be an integral part of the annual volume, since it may not be exported, it may be filed and ultimately bound with Volume 54 if so desired.

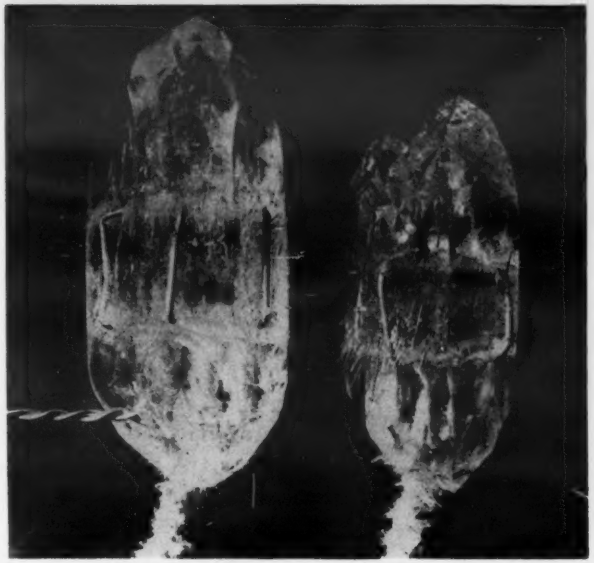

FIGURE 1. TWO BASAL PLATES 5 MM. THICK UPON WHICH QUARTZ HAS GROWN BOTH ABOVE AND BELOW

Dimensions of original plates are indicated by partly covered silver wires. According to Spezia, the fine, acicular crystals adhering to the large crystals formed during rapid cooling of apparatus at end of experiment. It is estimated from Spezia's text that about 14 mm. of quartz was added to the plate shown on the left.

FIGURE 2. QUARTZ CRYSTAL TWINNED AFTER THE JAPANESE LAW

Before being placed in the bomb.

FIGURE 3. SAME TWINNED QUARTZ CRYSTAL SHOWN IN FIGURE 2 AFTER THE ARTIFICIAL ACCRETION OF QUARTZ

Small prismatic stub extending horizontally toward the right is estimated to have grown about 5 mm. according to data by Spezia (1908).

ARTIFICIAL GROWTH OF QUARTZ ON SEED CRYSTALS

Produced in a silver-lined bomb by Giorgio Spezia.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 54, SUPPL. 1, 1 PL., 1 FIG. APRIL 1, 1943

RECORDED EXPERIMENTS IN THE PRODUCTION OF QUARTZ

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INTRODUCTION

In the last 100 years at least 30 investigators have recorded the production of quartz in the laboratory. Crystals formed have varied from a length of a few microns to 8 millimeters, and the length of natural

crystals used as "seed" crystals has been increased by 14 mm. The duration of experiments has ranged from 3 hours to 8 years, and the temperature of formation of the quartz from room temperature to 870° C. Pressures have varied widely, although in many cases the pressure of formation has not been reported. In spite of a number of obvious inaccuracies in descriptive data, little doubt exists that quartz has been produced in many experiments, as indicated by the crystal form, optical data, and physical properties of the product reported. Unfortunately, procedure has not been uniform; enumeration of the identifying properties of the quartz has frequently been omitted, and in a number of cases, where the grains have been only a few hundredths of a millimeter long and few confirming tests have been made, there may be doubt that quartz was actually formed. Many of the accounts fail to indicate whether the quartz produced had good crystal form.

Regardless of inadequacies in the literature, a review of all such experiments must precede any study of the conditions under which quartz forms either in the laboratory or in nature. It is hoped that such a record as is here given will stimulate additional laboratories to undertake studies in the synthesis of sizable quartz crystals. Our chief regret is that this paper must of necessity appear without a description of accompanying experimental work. The importance of conveniently grown crystals of useful size is so generally recognized that further comment is unnecessary.

Annotated bibliographies on the artificial formation of silicates have been prepared by Niggli and Morey (1913) and by Morey and Ingerson (1937). Many of the papers concerned with the production of quartz that are here included have been cited in these two papers. The present review attempts to classify, bring up to date, and correlate the recorded experiments concerning quartz alone.

CHRONOLOGICAL LIST OF EXPERIMENTS IN THE PRODUCTION OF QUARTZ

The methods of synthesis employed, the observer, and the year in which the experiment was attempted are as follows:

Page

Evaporation of a silica-bearing, aqueous solution at atmospheric pressure.
Before 1845—Bergmann; Siegling 4
1927—Rao 4
Crystallization without water at atmospheric pressure.
At low temperature by devitrification of a glass.
1872—Benrath 4
At high temperature from dry melts.
1849—Daubrée 4
1878—Hautefeuille 5
1880—Marsden 5
1881—Hautefeuille and Margottet 7

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1893—Morozewicz 5
1899—Bauer 6
1904—Lenarčič 6
1906—Day and Shepherd 7
1913—Fenner 5, 6
1918—Wilson (Observed to form in glass furnace) 7
1930—Bowen, Schairer, and Willems 8
1936—Trostel 8
1939—Kracek 8
Crystallization from chloride vapors.
1854—Daubrée 9
1908—Brun 9
Crystallization in sealed glass containers at moderate temperature and pressure.
1862—Déville 9
1873—1881—1887—Chrustschoff 9
Crystallization in sealed metal containers at high temperature and pressure.
1845—Schafhäutl 10
1851—Sénarmont 10
1872—Maschke 13
1879—Daubrée 10, 11, 12
1879—1881—Friedel and Sarasin 18
1881—Chrustschoff 13
1883—Friedel and Sarasin 14
1889—Bruhns 11
1891—Friedel 14
1891—Chrustschoff 10
1895—Chrustschoff 18
1896—Friedel 14
1898—Spezia 14
1902—Baur 17
1905—Spezia 14
1906—Königsberger and Müller 11, 12
1906—Allen, Wright, and Clement 17
1907—Day, Allen, Shepherd, White, and Wright 17
1908—Spezia 16
1911—Baur 17
1912—1913—Fenner 13, 19
1914—Schlaepfer and Niggli 18
1917—Morey and Fenner 19
1918—Müller and Königsberger 20
1925—Weil 14
1927—Ipatieff and Mouromtseff 13
1927—Morey and Bowen 12
1939—1940—Michel-Lévy, Albert and Wyart (Detonation method) 19
1940—Ingerson and Morey 11
1941—Geophysical Laboratory 22
Crystallization in the presence of superheated steam at atmospheric pressure.
1906—Quensel 20

ACKNOWLEDGMENTS

This paper, in manuscript form, was submitted to various investigators interested in the problem of the synthesis of quartz. The writers wish to take this occasion to express their appreciation of the wholehearted co-operative response which has been received. Not only suggestions and comment have been furnished which have been used to advantage, but discussion of work in progress and yet to be published has been freely offered. Dr. C. N. Fenner and Dr. Earl Ingerson have been most helpful in describing the excellent studies of the Geophysical

Laboratory. Dr. R. R. Williams, Dr. S. O. Morgan, Dr. W. L. Bond, Dr. A. N. Holden, and Dr. M. D. Rigterink of the Bell Laboratories have most considerately gone over the work of their organization on a number of occasions. Dr. R. B. Sosman of the U. S. Steel Research Laboratory has drawn freely upon his long experience in the problems of silica. Dr. N. L. Bowen of the University of Chicago has kindly offered comments.

SYSTEMATIC REVIEW OF EXPERIMENTS

EVAPORATION OF SILICA-BEARING, AQUEOUS SOLUTION AT ATMOSPHERIC PRESSURE

Schafhäutl (1845) mentions earlier experiments by Bergmann and also by Siegling in which quartz crystals were said to have been produced from silica solutions allowed to evaporate at room temperature and atmospheric pressure for a period of years. No reference is given to any publication by these authors. Maschke (1855), performing a similar experiment, allowed a 2.49 per cent solution of silica to evaporate freely and obtained only tough transparent plates having the properties of opal. He believed that such plates had been mistaken for quartz by previous investigators.

Rao (1927) produced quartz in the course of an experiment on the formation of bauxite. A 3 per cent solution of sulfuric acid was allowed to drop at the rate of three drops a minute onto granulated basalt (ten mesh) in a hard glass test tube. The experiment was continued for 6 months at 32°-52° C. Rao reports abundant quartz, augite in different stages of alteration, and no feldspar. Augite partly altered to quartz and often looking spongelike was seen, but the quartz was too abundant to be derived entirely from augite and could also have been derived from the feldspar. Rao did not describe the form in which the quartz occurred.

CRYSTALLIZATION WITHOUT WATER AT ATMOSPHERIC PRESSURE

Low temperature, by devitrification of a glass.—Benrath (1872) devitrified a soda-lime glass by gentle heating for a long time, producing small prismatic crystals of quartz and a feldspar near oligoclase. Day, Allen, Shepherd, White, and Wright (1907, p. 225) state, "Quartz cannot be formed by direct crystallization of silica glass since at the lower temperature at which it is stable the viscosity of the glass is too great to permit adequate molecular movement for crystallization."

High temperature, from dry melts.—Daubrée (1849) heated silicon chloride to white heat in a porcelain tube (not altogether satisfactory), in an earthenware retort, and in a crucible. In two of these experiments he obtained a deposit of silica with a "vitreous fracture," the mammillary

surface of which showed here and there very small crystalline faces including triangular faces like those of quartz.

Hautefeuille (1878) found that quartz in double hexagonal pyramids formed when amorphous silica was heated with sodium or lithium tungstate for several hundred hours at 800°-950° C. The pyramids were more acute with the lithium tungstate than with the sodium tungstate.

Marsden (1880) produced hexagonal, colorless, transparent prisms in a melt of metallic silver and amorphous carbon in a silica crucible which was kept above the melting point of silver for "a number of hours" and subsequently cooled gradually for 12-14 hours. An examination of thin sections of heated and unheated crucibles showed that the glaze which was vitreous silica before heating "had become one mass of little crystals in the form of hexagonal prisms." Dendritic laminae with roughly hexagonal outlines were also formed, varying from light yellow to dark brown or black. These were more numerous than the prisms, and Marsden at first mistook them for graphite. He suggested that the form of these siliceous plates, not known in natural quartz crystals, might have resulted from the silica being in true solution in the silver. According to Sosman (1927, p. 55) these "leaf crystals" were probably tridymite.

Morozewicz (1893) produced quartz crystals about 0.005 millimeter long from a melt of the composition of liparite (rhyolite) with 1 per cent tungstic acid which was kept at a temperature between the melting point of leucite (variously given in the literature as 1275°-1430° C.) and that of nepheline (variously given in the literature as 1059°-1526° C.) for 31 days. The same mixture without tungstic acid gave tridymite crystals in a glassy groundmass after 2 weeks, whereas, with the addition of 5 per cent tungstic acid, no trace of crystallization was observed even after 10 days. The 1 per cent tungstic acid may have been responsible for the formation of the quartz, but since the duration of each experiment differed from that of the others it is impossible to decide on the greater importance of time or the catalyst as a factor in crystallization.

Fenner (1913), investigating the stability relations of the silica minerals, found that the inversions were excessively sluggish unless a flux was present. "By heating either quartz or amorphous silica without a flux, cristobalite will be obtained at temperatures well below its field of stability." (p. 343) Concerning a fluxing or solution agent Day, Allen, Shepherd, White, and Wright (1907, p. 225) reported:

"The chief function of such a flux is to increase molecular mobility of the crystallizing mass so as to give it greater freedom and power of crystallization. From many experiments in our laboratory it has been observed that crystals produced

directly from silicate melts were commonly small and badly formed, while the addition of small percents of a second substance markedly increased the size and quality of the crystals even when the added substance hardened as glass and apparently did not enter into the composition of the crystals."

Fenner (1913) tried sodium tungstate, potassium and lithium chlorides, boric acid, and salt of phosphorus, finally choosing sodium tungstate ($Na_2WO_4 \cdot 2H_2O$) because it melts at a low temperature and does not volatilize at a high temperature. With this flux he heated silica glass at 863°-875° for 11 hours, 20 minutes; artificial tridymite at 854°-864° for 73½ hours; chalcedony at 750°-850° (time not given); "ground quartz" at 865°-877° for 24 hours.[1] In every case the product was mostly tridymite, but some quartz was produced in each case in rounded or distorted crystals or in simple prisms and pyramids, seldom exceeding 0.1 mm. in length. In one instance, when amorphous, precipitated silica was heated with a large excess of sodium tungstate over a Bunsen burner for 43 hours, quartz, tridymite, and cristobalite, all in good crystals, were formed in the same melt. In the crystallization of silica from silica glass or precipitated silica with the sodium tungstate flux, heated

"for a number of hours at 800°-900°, not quartz, but tridymite was first obtained. It was only after much longer heating that quartz crystals began to appear although this is the stable form at that temperature. It seems that in the passage from the amorphous condition to quartz, the whole is first converted into the intermediate form tridymite and only secondarily into quartz." (p. 339).

According to Fenner, "neither quartz nor tridymite has been formed under any conditions in the absence of a solvent."

Bauer (1899) considered the presence of tungstate responsible for the formation of quartz from a melt which he heated to 1100°. He fused orthoclase, albite, mica, hornblende, sodium chloride, sodium tungstate, boric acid and sodium phosphate in a porcelain crucible.

Lenarčič (1904) fused 5 grams of chlorite schist powder, 1 gram of calcium chloride, and 0.5 gram of potassium fluoride in a crucible by heating only 1 hour (temperature not given). This mixture was held fluid for half an hour, then cooled to 1020° and kept at that temperature for 5 hours, after which it was slowly cooled for 6 hours. Several minerals formed, among them quartz.

A similarly conducted experiment (Lenarčič, 1904) with 5 grams of powdered biotite gneiss and 0.7 gram of potassium tungstate produced a plagioclase-pyroxene rock not unlike basalt, containing occasional quartz plates.

[1] In many accounts the writer fails to indicate whether the temperatures are given in degrees Centigrade or Fahrenheit. No attempt has been made in this paper to correct the omission, although it is probable that in most cases degrees centigrade are implied.

With the addition of 0.5 gram of magnesium fluoride to 5 grams of biotite gneiss and 0.5 gram of potassium tungstate a rhyolitic glass formed, also containing quartz in small plates. This last experiment was repeated, using a mixture of the same chemical composition as the biotite gneiss instead of the natural rock and a little less quartz was formed.

Five grams of a mixture of the chemical composition of a mica schist were fused with 1 gram of molybdic acid (Lenarčič, 1904), 1 gram of lithium chloride, and 1 gram of magnesium fluoride; plates of quartz (*Quarzdurchschnitte*) were visible in thin section. When 5 grams of amphibolite schist powder were fused with 0.5 gram of tungstic acid and 0.7 gram of potassium tungstate, no quartz was formed. In almost all other experiments where the form of artificial quartz is described, it is reported as in doubly terminated prisms or in bipyramids. Siliceous plates have, in other cases, been identified as tridymite. In consideration of the temperatures employed by Lenarčič it seems possible that tridymite, not quartz, was formed.

In 1906 Day and Shepherd (1906a; 1906b) obtained quartz crystals from glass by the use of a mixture of 80 per cent potassium chloride with 20 per cent lithium chloride "at all temperatures below 760°."

Hautefeuille and Margottet (1881) state that hexagonal plates of tridymite form at bright red heat when lithium chloride is used as a flux, while below the fusion point of silver (960° C.) quartz forms in prisms terminated by double pyramids. No description is given of the apparatus or the form of silica used.

In 1918 a glass furnace in Fife burst as a result of the solvent action of the glass on the brick, and the 70 tons of molten manganese glass which constituted the charge cooled gradually for 5 days. Wilson (1918) reports that in the chilled glass he found quartz in short bipyramids with prism faces absent or small. From another furnace a pot that had contained lead glass (50 per cent lead) was coated with a 4-inch white layer in which tridymite and a little quartz were found.

In a number of experiments on silicate melts conducted at the Geophysical Laboratory of the Carnegie Institution in Washington quartz has been formed at atmospheric pressure. In these experiments the original constituents have been crystalline, formed by devitrification of a glass of the desired composition. The charge has been wrapped in platinum foil and heated in an electric furnace long enough to insure equilibrium. The melt has then been dropped into a dish of mercury at room temperature, and the resulting solid mass has been examined under the polarizing microscope.

Bowen, Schairer, and Willems (1930), investigating the system Na_2SiO_3-Fe_2O_3-SiO_2, found quartz formed from melts as follows:

Weight per cent			
Fe_2O_3	Na_2O	SiO_2	Temp. ° C.
10.00	18.46	71.54	882
10.00	18.46	71.54	893
7.50	21.61	70.89	800
12.50	17.95	69.55	845
12.50	17.95	69.55	850
14.00	16.51	69.49	860
.....	25.60	74.40	816
1.25	25.28	73.47	785
5.00	22.20	72.80	835
5.00	22.20	72.80	837
11.75	18.10	70.15	861
11.75	18.10	70.15	865
12.50	17.95	69.55	845
12.50	17.95	69.55	850
14.00	16.51	69.49	860

On the diagram drawn from these data the highest temperature of formation of quartz is 870° C.

Trostel (1936) fused 100-mesh glass sand at 1760° C., crushed the resulting glass to grains that would pass a 6-mesh sieve but were retained on a 20-mesh sieve, and held these at 1300° for 4 hours. At the end of this time the charge was water-quenched and examined under the polarizing microscope. Quartz "crystals" were found in the glass, occurring as "rounded, irregular, or anhedral-shaped inclusions, varying in diameter from 0.0036 to 0.0180 mm." A commercial form of vitreous silica, treated in the same way, did not give quartz. Trostel points out that extrapolation of the vapor pressure curve of quartz above that of tridymite would indicate the possibility of the formation of quartz as a metastable phase at 1300°.

Kracek (1939), working on the system Na_2SiO_3-Li_2SiO_3-SiO_2, found that "the rate of dissolution of quartz increases rapidly with even small additions of lithium and, within the system, the reactivities of quartz and tridymite are sufficiently high to permit their mutual conversion (quartz⇆tridymite) near the inversion temperature in 24 hours." Kracek produced quartz from melts having the following compositions and temperatures:

Weight per cent			
Na_2O	Li_2O	SiO_2	Temp. ° C.
14.98	7.69	77.33	805
14.98	7.69	77.33	795
19.98	3.50	76.52	829
19.98	5.00	75.02	688
19.98	5.00	75.02	642
21.99	2.00	76.01	827
21.96	3.00	75.04	739
22.00	3.70	74.30	686
15.29	7.14	77.57	828

Lithium compounds have been used by Hautefeuille (1878), Hautefeuille and Margottet (1881), Day and Shepherd (1906), and Kracek (1939).

CRYSTALLIZATION FROM CHLORIDE VAPORS

Daubrée (1854) passed silicon chloride vapor over "lime," "magnesia," "alumina," and "glucina" (beryllia) and formed quartz crystals in hexagonal prisms and pyramids. He does not mention confining pressure as a condition of the experiment.

Brun (1908) passed mixed sodium and potassium chloride vapors over masses of fused silica at 700°—750° out of contact with air. After 40 hours fine fibrous quartz had been produced, and after 3 or 4 days the nearly parallel fibers had united to form optically homogeneous crystals up to 3 millimeters long.

CRYSTALLIZATION IN SEALED GLASS CONTAINERS AT MODERATE TEMPERATURE AND PRESSURE

Déville (1862) produced "crystalline silica" by heating mixtures of potassium silicate and sodium aluminate solutions in a sealed glass tube. The solutions were of such proportions that the ratio of oxygen in the silicate stood to that in the aluminate as 2 to 1. At 170° levynite crystallized in small hexagonal tabular crystals. At a "higher temperature" "crystalline silica" formed in granules ("un sable").

Chrustschoff (1873, republished in 1881) kept dialyzed silica and water in sealed glass tubes for 10-14 days at 150° C. and hydrous silica at 250° C. for the same length of time. Both substances were dried at 120°, and the products in both cases were microscopic quartz crystals with hexagonal prisms and pyramids distinguishable at 300× magnification and a granular, noncrystalline material.

In 1887 Chrustschoff heated four glass spheres, half filled with aqueous dialyzed silica at 250-320° C., "almost every day for one to three hours over a period of six months." All the containers broke during the experiment, but in the one that lasted longest well-formed quartz crystals were found, the largest 8 millimeters long and 3 millimeters thick. Allen, Wright, and Clement (1906) regret that Chrustschoff did not give further information concerning the glass vessels capable of undergoing such an experiment.

In each case Chrustschoff used dialyzed silica, but the method of preparation of the silica differed. In the experiment in which the 8-millimeter crystal was formed (1887) the silica was obtained by dialysis of "a solution of 112 grams of potassium silicate (absolutely pure potassium hydroxide saturated with amorphous silica prepared from silicon fluoride) in a liter of distilled water to which was added 68g. concentrated hydrochloric acid."

In the less successful experiment (1873), however, the silica was obtained by dialysis of a dilute solution of sodium waterglass with a small amount of hydrochloric acid.

Probably traces of fluorine or a fluoride were responsible for the large size of the crystal.

CRYSTALLIZATION IN SEALED METAL CONTAINERS AT HIGH TEMPERATURE AND PRESSURE

Most of the experiments resulting in the production of quartz crystals have been performed in steel bombs or thick-walled tubes, usually lined with copper, silver, gold, platinum, or platiniridium, at a wide range of pressures up to 3000 kilograms per square centimeter. Experiments of this type are so numerous that discussion of them will be facilitated by grouping comparable experiments rather than considering them in chronological order.

The earliest published work on the artificial production of quartz is that of Schafhäutl (1845) who claimed to have produced microscopic crystals by 8 days' "evaporation" of an aqueous solution of freshly precipitated silica in an autoclave ("Papin digester"). The pressure in such a container probably did not exceed 2 atmospheres, and the temperature was probably not greater than 120° C.

In 1891, in an experiment designed to produce hornblende, Chrustschoff used evacuated 25cc. glass tubes in sealed iron tubes containing alumina, ferric and ferrous oxides, calcium-, sodium-, potassium-, and magnesium hydroxides and dialyzed silica. The containers were enclosed in sealed iron tubes and heated at 550° C. "for three months with many interruptions." The products were hornblende, quartz (in crystals 0.2 mm. long), and probably analcime and adularia.

Sénarmont (1851) performed experiments in an evacuated sealed glass tube enclosed in a sealed gun barrel which was almost filled with water. He placed in the glass tube a solution of bicarbonate of soda, 1 or 2 drops of alkaline silicate, and an excess of orpiment or realgar. When heated (temperature not given) the sulfides replaced the carbonic acid without acting themselves on the gelatinous silica. In each experiment only about 1 or 2 milligrams of fine quartz were obtained, most of which was granular, but some of which formed doubly-terminated hexagonal prisms. In all cases the grains "affected polarized light."

Daubrée (1879) heated water in a sealed glass tube in a sealed iron tube at about 400° "for a few days" and produced well-formed quartz crystals. Barus (1900), however, heated 210 grams of finely powdered soft glass (composition not given) and 50 grams of water at 210° for 12 hours in a steel bomb and produced only a hard, opalescent, homogeneous substance.

Daubrée (1879, p. 174) heated water from the hot springs of Plombières in a sealed glass tube within a sealed steel tube for 2 days at a temperature of "at least 400°" (p. 157) ("au moins de 400°"). . . . "Twenty to thirty cc." of this water "which is comparatively rich in silicates of potassium and sodium" was rapidly evaporated to about 1 cc. which was used in the experiment. After 2 days "the walls of the tube" were covered with chalcedony and quartz crystals. Since the glass was "altered" only at its surface Daubrée believed that the siliceous deposits must have come "almost entirely" from the decomposition of the alkaline silicate dissolved in the water.

Bruhns (1889) heated potash glass and a water solution of ammonium fluoride in a platinum tube in a steel bomb. After 50 hours at 300° quartz crystals were produced. Sodium glass or free silica gave quartz crystals after 10 hours at 300°.

Ingerson and Morey (1940) found 50 grams of quartz deposited near the top of a steel bomb in which 250 experiments had been performed at 3000 lbs. per sq. in. pressure and 600°-1050° C. The silica was reported as derived from the refractory tubes in the bomb and could not have been carried except through gaseous water.

Two experiments with glass in which the composition of the glass was known were those of Königsberger and Müller (1906) and Morey and Bowen (1927). Königsberger and Müller used glass of the following composition:

	Per cent		Per cent
SiO_2	69.21	MgO	0.52
Al_2O_3	2.48	CaO	9.84
Fe_2O_3	0.45	K_2O	1.98
Na_2O	14.91		

They used a platiniridium-lined steel bomb and describe the disadvantages of any less costly linings. The apparatus was equipped with a device for filtering the products of the reaction at high temperature and pressure within the bomb by rotation of the entire furnace.

Four experiments were performed, as follows:

1. 8g. crushed glass (2-5mm.), 60cc. dist. water, 2g. CO_2 60 hrs. at 360°, filtered 12 hrs. at 190°, cooled 8 hrs.
   Products: In filter tube: glass, double-ended qtz. xls. 0.05 mm. long.
   In main tube: Birefringent aggregate.
2. 8g. crushed glass, 60cc. water, 10 g. CO_2 heated as in (1).
   Products: In filter tube: No precipitate.
   In main tube: Quartz and chalcedony.
3. 8g. crushed glass, 60cc. water, 15cc CO_2 heated as in (1).
   Products: In filter tube: No precipitate.
   In main tube: Birefringent grains—quartz?
4. 10g. crushed glass, 50cc. water heated as in (1).
   Products: In filter tube: 2g. substance, chiefly quartz, in double-ended crystals, some 0.6 mm. long. Some with spindle-shaped liquid inclusions.

Morey and Bowen (1927) investigated the effect of water at high temperature and pressure on optical glasses of various compositions. In each case the glass and water were heated in a steel bomb for about 20 hours. The following summarizes their work:

SiO_2	B_2O_3	K_2O	Na_2O	CaO	BaO	ZnO	PbO	As_2O_5	T°.C.	Products
71.2	0.5	14.4	5.5	6.5		1.2		0.7	300	qtz. and amorph. subst.
									550	qtz. and needles
71.0		9.0	8.0	8.0	2.6			1.4	300	" "
									550	" "
71.5		5.6	7.1	10.4		4.3		1.1	300	" and amorph. subst.
									550	" and unknown
70.8	7.2	14.4	5.4	2.0				0.2	300	rhomb. and hexag. xls.
									550	qtz. and needles
65.8		1.4	14.8	2.6		1.6	13.2	0.6	300	qtz. and needles
									550	tridymite, cristobalite, qtz.
64.5		2.1	9.8	1.4			21.9	0.3	300	qtz. and two unknowns
									550	excellent quartz and one unknown
60.2		6.0	7.0				26.4	0.4	300	qtz. and two unknowns
									550	large qtz. xls.
45.8		4.6	4.0				45.0	0.6	300	excellent qtz. and hexag. plates, high index of ref.
									550	excellent hexag. uniax. xls. with high index (1.7+).

It is noteworthy that the best quartz crystals formed in glass with a high lead content. This gives added weight to Wilson's account of quartz formed in lead glass from the glass furnace at Fife (1918).

Daubrée (1879) and Königsberger and Müller (1906) used obsidian as a source of silica. Daubrée found that the action of superheated steam on obsidian (or on feldspar or pyroxene) in a sealed glass tube in a steel bomb caused it to become completely coated with quartz crystals. Königsberger and Müller used Lipari obsidian of the following composition:

	Per cent		Per cent
SiO_2	74.3	CaO	1.0
Al_2O_3	13.0	Na_2O	3.8
Fe_2O_3	2.6	K_2O	4.6
MgO	0.3	H_2O	0.3

Five experiments were performed, as follows:

1. 10g. obsidian, 60cc. distilled water. 48 hrs. at 320°, filtered 12 hrs., cooled 12 hrs. to 240°, room temperature 6 hrs.
   Products: No quartz.
2. 10g. obsidian, 50cc. distilled water, 15g. CO_2, treated as in (1).
   Products: No quartz.
3. 10g. obsidian, 60cc. distilled water, 3g. CO_2, 2g. $NaHCO_3$, treated as in (1).
   Products: In filter tube: A very small amount of quartz.
   On obsidian: A little chalcedony.

4. 10g. obsidian, 60cc. distilled water, 4g. $NaHCO_3$, treated as in (1).
Products: In filter tube: 0.3g. quartz.
Obsidian strongly attacked. Quartz and a dark amorphous, concretionary substance deposited on it.

5. 10g. obsidian, 40cc. of a solution of the composition of that found in cavities in quartz crystals, as follows:

	Per cent	
H_2O	85.	
CO_2	5.	
Na_2O	2.5	Heated 42 hrs. at 420°, filtered 12 hrs., cooled to 300° 12 hrs.
K, Li	1.5	
Ca	0.3	Products: In filter tube: Very small long qtz. xls.
CO_3	3.5	Precipitate: Crystals of aegerine augite.
Cl	1.5	
SO_4	0.7	

In the first four of the above experiments quartz was produced only when sodium bicarbonate was added. Fenner (1913, p. 356) states that

"quartz may be prepared without difficulty by heating either silica glass or amorphous precipitated silica with water and sodic carbonate in a silver-lined steel bomb at 400°-500° for 2-3 days. The relative proportions of materials need not be very exact; approximately the following were used in several experiments: 8cc. water, 2-3g. silica, 0.7g. crystallized sodic carbonate. Capacity of bomb 16cc."

A similar statement appears in an earlier paper (1912, p. 474). Chrustschoff (1881; 1895) and Ipatieff and Mouromtseff (1927) have performed experiments with silica and water alone. Chrustschoff (1881) heated hydrous dialyzed silica in a gold-lined cast steel bomb at 350° for 10-14 days. The product, dried at 180°, was a powder with a specific gravity of 2.25-2.30. With a magnification of 300× Chrustschoff observed hexagonal plates. Ipatieff and Mouromtseff (1927) made an emulsion of silicic acid gel and water, introduced it into a silver tube which was placed in a pressure bomb charged with hydrogen, and kept the whole at a temperature of 310°-320° and a pressure of 200 atmospheres for 30-40 hours. The product was washed with hydrochloric acid to eliminate oxides and examined under the microscope with weak magnification. Quartz with hexagonal prisms and pyramids was recognized.

Maschke (1872) used an evacuated sealed glass tube in a steel bomb. Fifty cc. of freshly prepared sodium hydroxide solution was strongly heated in a platinum dish with as much amorphous granular silica as would go into solution. This clear liquid was introduced into the glass container. After 24 hours' heating at 175°-185° the bomb was cooled to about 120° during 1-1½ hours, then allowed to cool completely. Hydrated silica and a small amount of a substance that "behaved in the same way as crystalline quartz" formed. When the temperature was increased to 190°-205° there was a marked increase in the amount

of quartz in the silica lumps; this was easily recognized in angular birefringent grains.

Weil (1925), using sodium silicate solution as the "mineralizer" with anhydrous precipitated silica heated in a steel tube at 650°-750° for 7-15 hours, found that with weaker solutions of sodium silicate cristobalite formed, whereas with stronger ones quartz was produced in acute pyramids, the prisms being reduced or absent. (*Cf.* Spezia's experiments, 1905.)

Friedel and Sarasin (1883) failed to produce quartz from a mixture of the composition of albite plus excess sodium silicate at 430°-500° in a sealed platinum tube. When various quantities of potassium chloride were added, however, some quartz crystals formed, with a quantity of an apparently isometric unknown substance. Other experiments under apparently identical conditions yielded no quartz, a fact which the authors were unable to explain. (*Compare* the effect of addition of potash by Michel-Lévy and Wyart (1939).)

Georges Friedel (1891) produced corundum and diaspore by the action of a soda solution on excess amorphous alumina in a copper-lined steel tube at 450°-500° for 6 to 10 hours. If the alumina contained a little silica, it was precipitated as quartz crystals. "It is strange that in the presence of a great excess of alumina, it was not albite which formed, but quartz, whereas with a smaller proportion of alumina the feldspar formed readily." (p. 9)

In 1896 Friedel investigated the action of a silica-bearing sodium hydroxide solution on muscovite in a platinum-lined steel tube and found that if the proportions of silica to soda in the solution were as 4:1 albite and quartz were formed. At the end of 2 or 3 days at about 500° a few sparse albite crystals and an abundant deposit of quartz in bi-pyramids, forming a veritable concretionary vein in the tube, had formed.

Spezia (1898) used sodium silicate in a variety of inadequately described experiments. In 1905, however, Spezia (1905a) described in detail additional growth on natural crystals in a sodium silicate solution in a silver-lined bomb. The bomb, whose inner dimensions were about 12 by 2 inches, was placed in a vertical position, the upper part being kept at 320°-350° and the lower part at 165°-180° (presumably C.). The bomb was filled with a 2 per cent solution of $Na_2SiO_3 \cdot 8H_2O$ and a silver-wire basket of quartz fragments without crystal faces was placed in the upper part of the bomb. One perfect crystal was set on top of the fragments. The basket was so placed that its upper part was above the annular gas pipe used for heating the bomb and its lower part was

below the heating ring. One centimeter below the bottom of this container three quartz crystals were fastened on a silver wire.

Heating the upper part of the bomb and cooling the lower part allowed slow diffusion of the hot, silica-bearing solutions to the cooler region where deposition would take place.

At the end of 6 months, the quartz fragments in the upper part of the container, as well as the crystal on top of them, showed signs of extensive corrosion. The fragments at the heating ring and below were cemented together and difficult to separate. Striated crystal faces had developed on the free surfaces of the fragments, and several of the wires of the basket were imbedded in the quartz and partly covered. One of the seed crystals had grown slightly, but in this experiment they were apparently below the region of maximum deposition. A loose, powderlike deposit of microscopic crystals was found in the bottom of the bomb, and Spezia suggests that these were probably formed freely in the solution at the region of maximum crystal growth, settling through the cooler portion of the solution to the bottom of the bomb.

The pressure, calculated by Spezia to be not greater than 150 atmospheres, was approximately uniform throughout. Spezia points out that if it had been an important factor assisting solution of the quartz above the heating ring it should have impeded deposition below the heating ring. If, on the other hand, it was an important factor in deposition, it should have impeded solution. He therefore concludes that the sole effect of the pressure was to keep the solvent in the liquid state and that solution and deposition were accomplished as a result of difference in temperature.

According to Spezia (1905b) the corrosion of the quartz in the upper part of the silver-wire container resulted in the formation of sodium polysilicate, the solution of which, being heavier, descended to the cooler part of the bomb where the excess silica was deposited as quartz.

It is noteworthy that the region of deposition was within a few centimeters of the region of solution, well above that portion of the bomb in which the temperature varied between 225° and 237°, as shown in an accompanying diagram of the apparatus.

In a second, similar experiment, lasting 199 days (1905b), the seed crystals were more advantageously placed. Two of these were plates ½ cm. thick[2], cut normal to the optic axis. At the end of the experiment one of these had increased in length to about 2½ cm. A photograph of these plates as they appeared at the end of the experiment (Pl. 1, fig. 1) shows that the silver wires holding the original plates became

[2] This and subsequent measurements were made from Spezia's photographs. No other record of the size of the crystals is given.

deeply buried by new growth except on the prism faces; the diameter of the crystals remained virtually unchanged. In this experiment, according to Spezia (1905b), "above 326° was the temperature of the region of solution and below 326° was the temperature of the region of deposition."

Although the silver wire did not prevent the growth of the quartz there was some evidence that its presence produced turbidity in the newly crystallized material. In subsequent experiments, therefore (1908), the prepared prisms were secured with a girdle of silver wire, laid in notches cut around the prism normal to the optic axis. Two such prisms, approximately 5 mm. long, were placed in the bomb with an aqueous solution containing 1.9 per cent Na_2SiO_3 and 12.7 per cent NaCl. The source of the silica and the temperature of the bomb were the same as in the earlier experiments. One of the prisms was approximately 11 mm. in diameter, and the other approximately 6 mm. in diameter, so that the area of the cut surface of one was approximately three times the area of the cut surface of the other. After five months the larger crystal was about 19 mm. long, somewhat turbid, and with poorly developed rhombohedral faces; but the smaller crystal was about 13 mm. long, with well-developed faces, the newly grown part of it being perfectly transparent.

Spezia interpreted the results as indicating that the amount of growth was roughly proportional to the area of the cut surface from which the growth occurred and suggested that the difference in turbidity in the two crystals resulted from the different rates of growth. In the same experiment Spezia produced additional growth on a Japanese twin with two broken terminations (Pl. 1, fig. 2). The added quartz was water clear, although the original specimen had been somewhat turbid (Pl. 1, fig. 3). Spezia noted that the presence of the sodium chloride in the solution produced crystals that were more transparent, had striated prism faces, and showed greater diversity in size between the two sets of rhombohedral faces. "The presence of the NaCl caused the limpidity of the crystals to be identical with that of the better crystals found in the geodes in the Carrara marble." (p. 97)

In order to investigate the effect of the movement of the solutions, an experiment was performed with two prepared prisms, one with its optic axis vertical and the other with its optic axis horizontal. After 5 months the horizontal prism, having its cut face parallel to the length of the bomb, had increased in weight 78.65 per cent, while the vertical prism, having its cut face normal to the long direction of the bomb, had increased in weight 96.98 per cent. The solution used in this experiment contained 11.3 per cent NaCl and 1.2 per cent Na_2SiO_3.

Spezia's results command special attention, not only because of the quality and volume of quartz produced, but also because the system involved is simple enough to admit of systematic physical-chemical investigation.

Allen, Wright, and Clement (1906, p. 406) kept large lumps of sodium silicate with magnesium ammonium chloride in a steel bomb at 400°-450° for 3 days and produced clear, doubly terminated quartz crystals, sharply defined crystallographically, the largest of which measured nearly 3 millimeters in length. In a later paper, Day, Allen, Shepherd, White, and Wright (1907, p. 224) describe the same experiment, giving the maximum length of the crystals at 2 mm.

Baur (1902) performed a series of experiments in a sealed steel tube on the conditions of formation of orthoclase and albite. In each experiment the temperature was held at 520° for 4-6 hours, cooled to about 450° gradually over a period of 2-3 hours, then the heat was turned off and the furnace allowed to cool to room temperature. The amorphous silica used was produced by the introduction of silicon fluoride in water. If minute amounts of fluoride were retained with the silica they might have influenced the crystallization of the quartz. (*Compare* Schlaepfer and Niggli, 1914.) Six experiments were performed with soda, as follows:

						Products
1.	13g.	SiO_2	4.5g.	$NaAlO_2$,	12cc. H_2O	quartz only
2.	5g.	"	2.3g.	"	" "	quartz only
3.	"	"	4.3g.	"	" "	quartz, tridymite, sodic albite
4.	"	"	10. g.	"	" "	albite only
5.	"	"	15. g.	"	" "	unidentified mineral similar to nepheline
6.	"	"	25. g.	"	" "	unidentified mineral similar to nepheline

With the same procedure, Baur performed similar experiments with potash, as follows:

						Products
1.	5g.	SiO_2,	2g.	AlO_2K,	12cc. H_2O	quartz in well-developed forms and an unidentified radial mineral
2.	"	"	3.3g.	"	" "	quartz, orthoclase in globules
3.	"	"	5. g.	"	" "	quartz, orthoclase predominating, some in globules, some xls.
4.	"	"	15. g.	"	" "	orthoclase, well formed
5.	"	"	25. g.	"	" "	unidentified needles

Baur used an unlined steel cylinder with a screw top. Contrary to the opinion of Königsberger and Müller, he maintained that a lining of precious metal was not necessary. (*Compare* Michel-Lévy and Wyart, 1939.)

In 1911 Baur reported further experiments with soda or potash, silica, alumina, lime, and water, performed in a similar bomb but at lower temperatures and for longer periods of time. In these experiments the

bombs were in the furnace 20-24 hours. The temperature in the furnace was 350°-400°, but it took 8 hours for the bombs to become heated to this temperature. The following summarizes these experiments:

0.3g. $KAlO_2$, 0.7g. SiO_2, 1.8g. H_2O at 350° amorphous; 450°,	quartz and "kaliumfujasit"
1.0g. $Ca(AlO)_2$, 10.4g. waterglass, 0.6g. SiO_2, 8.6g. H_2O 450°	quartz and pectolite
1.2g. $CaCO_3$, 7.8g. water glass, 0.4g. $NaAlO_2$, 0.3g. SiO_2, 5.5g. H_2O 450°	quartz, gyrolite, pectolite
1.0g. $CaCO_3$, 7.8g. waterglass, 5.4g. H_2O 450°	quartz and pectolite
0.6g. $NaAlO_2$, 0.7g. SiO_2, 3.1g. H_2O 450°	quartz and albite

Schlaepfer and Niggli (1914), working with a silver crucible in a silver-lined autoclave at 470° for 1-2 days, found that a mixture of 1.30g. of silica (Merck), 0.105g. of potassium hydroxide, and 2cc. of water produced poorly formed crystals of quartz, whereas good crystals resulted in nearly all cases when a mixture of 1.301 grams of silica (precipitated from SiF_4 and possibly containing traces of HF), 0.105-0.378 grams of potassium hydroxide, and 2cc. of water were used. Addition of small amounts of alumina or lime did not prevent the formation of the quartz. Probably the small amounts of HF present were responsible for the formation of better quartz crystals.

Friedel and Sarasin (1879; 1881) also conducted experiments for the purpose of producing orthoclase. When "gelatinous silica, alumina and potash solution" were heated below dark red heat for 16-30 hours in a steel tube lined with copper or platinum quartz was formed in almost all the experiments and "sometimes in crystals big enough to be measured by the Wollaston goniometer." The two accounts of these experiments differ in detail.

Chrustschoff (1895) heated gelatinous or colloidal silica, gelatinous or colloidal alumina, and potash in a sealed platinum vessel at 300°-400° and produced quartz, tridymite, orthoclase, and corondum. When soda was substituted for the potash, albite and corundum formed, but no quartz. (*Compare* Friedel, 1891.) When Chrustschoff heated colloidal silica, colloidal alumina, colloidal iron oxide, colloidal hydrous iron oxide, calcium hydroxide, magnesium hydroxide, potash, and soda to 550° quartz, hornblende, diopside, orthoclase, and a hydrous silicate formed. Colloidal silica alone, heated to 250° or to 360°, gave quartz. Chrustschoff prepared a noncolloidal, water-soluble form of silica from silicon fluoride gas with the aid of dry boric acid and ammonia. A thick solution of this was placed in a platinum tube with one drop of hydrofluoboric acid (HBF_4) and presumably the tube was sealed. At temperatures above 240° quartz was formed after 5 hours' heating. At temperatures of 180°-228° hard isometric crystals were formed. At

240°-300° quartz was obtained, and at 310°-360° the predominant product was tridymite, but a little quartz persisted.

Fenner (1913, p. 358)

"placed in a bomb of 18cc. capacity 4g. amorphous, precipitated silica; 3cc. hydrofluoboric acid, made by saturating a 40 percent solution of HF with B_2O_3; and 3cc. water. These were heated 22 hours at 350-380°. The product was mostly unchanged amorphous silica, with which there were a few small, but perfectly formed, crystals of quartz."

Morey and Fenner (1917) investigated the system H_2O-K_2SiO_3, using a gold crucible in a bomb, quick-chilling the bomb in cold water and examining the products with the aid of the polarizing microscope.

"The primary effect of pressure in systems of this type is to enable the concentration of the water to be kept at the required value. Such an effect of pressure is of great importance, as the concentration of the volatile component is a function of the pressure, and is to be sharply differentiated from the effects due to uniform hydrostatic pressure, which are comparatively small" (p. 1186).

The following tabular summary includes the runs recorded by Morey and Fenner in which quartz was produced and a few in which it was not produced, for comparison.

SiO_2/K_2O	Total g. H_2O	g. H_2O in solution	T. °C.	Product
4.05	21.0	0.27	420	quartz
3.59	9.5	0.34	420	$KHSi_2O_5$
3.34	4.43	0.26	480	"
4.27	4.86	0.26	480	quartz
4.49	5.82	0.35	480	"
4.27	4.81	0.22	500	$KHSi_2O_5$
4.49	5.42	0.32	500	quartz
4.63	8.9	0.36	500	"
3.29	0.32	0.08	520	$K_2Si_2O_5$
4.49	3.82	0.21	520	quartz
4.49	1.52	0.16	600	"
4.72	10.5	0.22	600	"

Michel-Lévy and Wyart (1939) produced abundant quartz crystals 0.4 millimeter long (with prism and pyramid faces) by detonation of silica and hexogene with 0.7cc. distilled water and 0.10-0.20 gram of "eau de potasse", the products being kept at 545° for 6 days at a pressure greater than 3000 kilograms per square centimeter, whereas, in a second experiment, identical except for the omission of the potash, cristobalite was produced. Both the quartz and the cristobalite were identified by X-ray analysis. Here, apparently, the potash was essential to the formation of the quartz.

When the silver tube used in these experiments by Michel-Lévy and Wyart was replaced by a red copper one (1940), the following results

were observed: With vitreous silica, water, and hexogene at 450°-650° quartz was formed: when considerable quantities of potash were added quartz was still formed (at 540°, 545°, and 585°), but, when a "one per cent solution of potash" was detonated with vitreous silica, water, and hexogene, cristobalite resulted, even though the tube was kept at 540°-545° for 4-13 days following the detonation. A significant comparison may be made between the two experiments that were performed without potash, one in a silver tube, producing cristobalite, and the other in a red copper tube producing quartz. The copper must have entered into the reaction and facilitated molecular mobility so that the form of silica stable at the prevailing temperatures was able to form.

Müller and Königsberger (1918), continuing their previous research (1906) with the same type of apparatus, produced quartz from potash water glass, dilute solution of "$AlOK_3$", and CO_2. The following summarizes the experiments in which quartz was formed:

Composition, in grams					T. °C	Time, in hours			Precipitate	In filter tube
K_2O	Al_2O_3	SiO_2	CO_2	H_2O		Heating	Filtering	Cooling		
6.58	0.64	5.67	0.64	30	320	24	12	48	K neph. leucite	1 mg. qtz. (0.1 mm. long) Trace of calcite
1.06	0.34	3.8	5.0	40	310	3	24	48	0.5g. qtz. (.01 -.03mm. long), leucite.	.05g. qtz. (.01mm. long)
"	"	"	"	"	420 to 430	"	"	"	.05g. qtz. (.01 mm. long), Kfspr., leucite, zeolite.	.5g. qtz. K feldspar

CRYSTALLIZATION IN THE PRESENCE OF SUPERHEATED STEAM AT ATMOSPHERIC PRESSURE

Quensel (1906) conducted superheated steam into a crucible containing "74 parts oligoclase, 26 parts silica" and 6 per cent WO_3, kept at 800°-1000° for 6 hours in a gas furnace; quartz crystals 0.05 millimeter long formed, but the same experiment continued for 10 hours without the WO_3 yielded crystals so small as to be scarcely distinguishable when magnified 1000 times. Increase of the amount of WO_3 to 100 per cent gave no better crystals than with 6 per cent. When amorphous silica and 500

per cent sodium tungstate were similarly treated at 750°-800° for 4 hours only quartz glass resulted.

DISCUSSION

CONTROL OVER EXPERIMENTAL CONDITIONS

The degree of control necessary over the conditions of an experiment which results in quartz seems to vary with the nature of the experiment. In some instances the limiting conditions are narrowly defined, while in others wide variations are possible.

Fenner (1913), in describing the preparation of quartz "without difficulty" from silica glass or amorphous precipitated silica with water and sodium carbonate in a silver-lined steel bomb at 400°-500°, says that the relative proportions need not be very exact. He further states that in all his experiments with amorphous silica in alkaline solutions quartz was obtained. However, an examination of the tabular data given by Morey and Fenner (1917) and comparison with Figure 1 show that slight differences in the proportions of the original constituents in the vicinity of the eutectic composition lead to the formation of different products.

Friedel and Sarasin (1883) at first produced many quartz crystals in an experiment with sodium silicate, aluminum silicate, and potassium chloride in a bomb but later performed other experiments under apparently identical conditions which did not result in the formation of quartz crystals. Finally, Ipatieff and Mouromtseff (1927), who produced quartz from an emulsion of silicic acid gel and water in a bomb, reported that minute traces of impurities may considerably change the results of such an experiment.

TEMPERATURE

Experiments in sealed glass containers have been performed at temperatures ranging from 180° C. to 320° C.; in sealed metal containers (in some cases containing sealed glass vessels) from 200° C. to 1050° C.; and in open crucibles from 750° C. to 1100° C. Experiments prior to 1845 and inadequately described indicated that quartz has been formed at room temperature by the evaporation of aqueous solutions. However, such formation may hardly be regarded as established. If available data on size and crystal form are to be accepted, it does not appear that as a group the higher-temperature experiments have produced larger or better quartz crystals than the lower-temperature experiments.

PRESSURE

In experiments where water is used at elevated temperature in a sealed container the pressure is of course elevated as well. The primary effect of pressure under these conditions is to enable the concentration of the

water to be kept at the required value. The concentration of the volatile component becomes a function of the pressure.

The pressure in such systems is important only because it produces a high concentration of the volatile component. The effects as pressure per se are comparatively small, according to Morey (1917).

Spezia interpreted his remarkably successful experiments in which solution and deposition took place at different temperatures, but at the same pressure, as indicating the negligible effect of pressure on such reactions. Whether his results would have been improved if the pressure in the region of solution had differed from the pressure in the region of deposition is still to be determined.

Pressures involved in the experiments reviewed range from atmospheric pressure to about 42,000 lbs. per sq. in., obtained by Albert Michel-Lévy and Wyart in detonation experiments. Except for these experiments, the upper pressure limit at which quartz has been produced in a sealed bomb appears to be 15,000 lbs. per sq. in., obtained at the Geophysical Laboratory in Washington (L. H. Adams, 1941).

The different experiments involving water in sealed glass containers without confining pressure on the outer walls have necessarily been performed at moderate temperatures, because the breaking strength of the vessels is low, and the pressure exerted by water vapor rises rapidly with rising temperature. Chrustschoff (1887) claimed to have heated four glass spheres "half filled with aqueous silica" to 320° C. at which temperature three exploded, but the fourth remained intact. The flasks had walls 5mm. thick and a capacity of 250 cc. The vapor pressure of water at 320° C. is 1637.6 lbs. per sq. in.

SIZE OF CRYSTALS

Spezia increased the length of a natural crystal by about 14 mm. at a temperature below 320° C. and possibly as low as 145° C.

Chrustschoff recorded the length of his largest crystal formed at 250° C. as 8 mm. With these exceptions, all quartz crystals whose lengths have been recorded as 0.1 mm. or greater have been formed at 400° C. or higher. It is possible, however, that some of the other experiments performed at lower temperatures would have resulted in the formation of larger crystals had they been continued for a longer time. Fenner (1913) made tridymite from silica glass with $Na_2WO_4 \cdot 2H_2O$ at 800°-850°, but with much longer heating the tridymite was converted to quartz.

The following summarizes those experiments in which quartz crystals 0.5 mm. in length or larger have been reported.

Length	Observer	Date	Constituents	Apparatus	C. Temp.	Continuity	Time
0.5 mm.	Friedel and Sarasin	1879	H_2O, "potash," pptd. alumina, excess gelat. silica	steel tube	below dk. red 500°?	cont.	38 hrs.
2 mm.	Allen and Wright	1906	sodium metasil., magnesium ammon. chloride, water	steel bomb Pt cruc.	400°-450°	cont.	3 days
3 mm.	Brun	1908	masses of fused silica exposed to KCl and NaCl vapors out of contact with air		700°–750°	cont.	3–4 days
8 mm.	Chrustschoff	1887	aqueous dialyzed silica	glass spheres 200 cc. capacity	250°–300°	intermittent	6 mos.
14 mm. increase	Spezia	1905 and 1908	sodium silicate, water, quartz	silver-lined steel bomb	165°–350°	cont.	100 days, 5 mos.

TIME

The element of time has frequently been advanced to account for the failure of laboratory experiments in the synthesis of quartz to equal or even approach nature's product from the standpoint of size. If this assumption is true, the inability of laboratories to continue experiments for thousands or even hundreds of years would preclude the possibility of forming large crystals artificially. While it is theoretically possible that large crystals may require great time intervals, the requirement cannot be considered as established. In any event our knowledge is at present too meager for definite conclusions.

According to Spezia (1905a) the deposition of quartz in his experiments was "due to a slow and continuous chemical process in which, to obtain a noticeable effect, time must be a contributing factor." Further, in his experiments with two prisms of different diameter, Spezia showed that the crystal that grew more slowly was clearer and better-formed than the crystal that grew more rapidly. All other conditions of formation being the same, the more rapidly grown crystals were likely to show one rhombohedral face much larger than all the others.

If the supply of any of the necessary constituents is so limited as to become exhausted in a short time, further prolongation of the experiment cannot increase the size of the crystals formed, but if a continuous supply of such constituents can be maintained there appears to be no

theoretical limit to the size of the crystals that may be formed. L. C. Graton (1940, p. 263, footnote 69) quotes an unpublished paper by E. B. Dane, Jr., suggesting that laboratory experiments on vapor-phase transport might well be made in dynamic systems rather than in closed bombs. The difficulty of controlling the conditions of such an experiment is obviously great.

The few experiments in which crystals .5 mm. or longer have been formed indicate that time may be an important factor in determining the size of quartz crystals produced in the laboratory. On the other hand, since temperatures and constituents of these experiments have varied widely, factors other than time may have been more instrumental in determining the size of the crystals.

COMPOSITION

When constituents other than silica are present in a charge, the melting point of the mixture may be considerably lower than that of pure quartz. In such a case, if the mixture is heated to a high temperature, quartz crystals will not form unless the mixture is subsequently slowly cooled. This may be illustrated by reference to the binary system K_2SiO_3-SiO_2 as worked out by Morey (1917). If a mixture of potassium orthosilicate and silica with a silica-potash ratio of 4:1 is held at 550° C., no crystals of any sort appear, regardless of the duration of the experiment. At this ratio if the temperature dropped to 525° C. quartz would form. (*See* Figure 1.) However, with a silica-potash ratio of about 4.1-4.2:1, quartz crystallizes out at the higher temperature. In fact, for this ratio 550° C. would be a more favorable temperature than 525° C. The experiments of Baur and Chrustschoff illustrate improvement in the conditions of crystallization resulting from such increase in temperature.

Baur (1911) heated 0.3 gram $KAlO_2$+0.7 gram SiO_2+1.8 gram H_2O at 350° C. for 12-16 hours and obtained only amorphous material, whereas with a temperature of 450° C. quartz was one of the products.

Chrustschoff (1873) working with dialyzed silica (apparently with a little water) in a sealed glass tube formed only amorphous material after 10-14 days at 150° C. After 10-14 days at 250° C. small crystals had formed which were just discernible microscopically at 300× and could be identified as hexagonal prisms at 700×.

Chrustschoff (1895) heated a true solution of silica (not colloidal) prepared from dry boric acid, silicon fluoride gas, and ammonia in a platinum tube. To each charge he added one drop of hydrofluoboric acid (HBF_4). No quartz formed after 5 hours at 228°-235°, but after 5 hours at 240°-300° quartz formed, and at 310°-360° quartz and tridymite.

In both sets of experiments by Chrustschoff a temperature of about 250° was apparently the minimum at which quartz would form in the length of time allotted.

Apparently the higher the temperature, provided it does not exceed

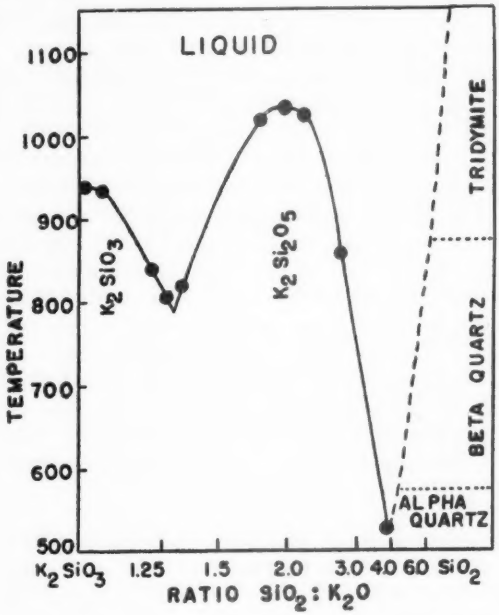

FIGURE 1.—*Variation of melting point with SiO_2:K_2O ratio on the 1-atmosphere isobar*

After Morey and Fenner (1917).

the liquidus temperature for the composition investigated, the more readily are quartz crystals formed. This is expectable inasmuch as higher temperature promotes mobility of the constituents and therefore facilitates crystallization.

It should also be pointed out that in the binary system just discussed, if the silica:potash ratio is less than 2:1 no quartz crystallizes out regardless of the temperature and duration of the experiment if equilibrium is maintained. In each case the crystalline product would be $K_2Si_2O_5$, K_2SiO_3, or both. When the ratio is between 2:1 and 4:1 the product is first $K_2Si_2O_5$, then quartz. With ratios above 4:1 the product is first

quartz, then $K_2Si_2O_5$, provided cooling is slow enough to maintain equilibrium.

Friedel (1896) produced a "veritable concretionary vein" of quartz in a sealed, platinum-lined, steel tube from the action of a silica-bearing sodium hydroxide solution on muscovite when the proportions were such that SiO_2:Na_2O in the solution was 4:1. When this proportion was 2:1 or 3:1, however, analcite and albite formed without quartz.

In a two- or three-component system the complete picture of the phase equilibrium of the system can be obtained. However, a review of the literature shows that many successful experiments have involved more than three components. Each of these experiments may be considered as an isolated point on a multi-component diagram. When the whole diagram is known, the optimum conditions of temperature and concentration for the formation of quartz crystals with any given set of constituents may be ascertained. Such knowledge is achieved only through long and painstaking experiment, and even when all the necessary information is assembled a system of five or more components is, to quote Morey (1936, p. 254), "almost inconceivable. . . . It is the human mind which places the limitation on our knowledge of the processes of Nature." Because a system of more than four components is so complex, other experimental routines than those followed in the preparation of the phase-relation diagrams may be preferred.

CONSTITUENTS USED

Various investigators have used many constituents. Some have been essential factors in the chemical reaction; others have behaved as catalysts. No one substance stands out as having been widely and exclusively used in the production of quartz crystals. Those used successfully by three or more investigators are selected as worthy of mention.

The conditions of the experiments in which potassium chloride, lithium chloride, and sodium tungstate have been used suggest that the presence of these compounds may be especially conducive to the formation of quartz crystals. This is particularly interesting since these constituents have not been widely used in experiments materials.

Silica.—Silica from the following sources was used in experiments resulting in the production of quartz:

gelatinous silica
amorphous silica
dialyzed silica
silica solution
precipitated silica
fused silica
$SiO_2.3H_2O$
chalcedony
quartz
tridymite
cristobalite
waterglass
sodium silicate
potassium silicate
potassium aluminum silicate

aluminum silicate	mica schist
silicon chloride	biotite gneiss
glass	orthoclase
lead glass	albite
sodium glass	oligoclase
calcium glass	muscovite
obsidian	hornblende
basalt	pyroxene

The terms used overlap in several cases, resulting in uncertainty as to the exact nature of the substance used. For example, "amorphous silica" may mean silica glass, opal, precipitated silica, or gelatinous silica. "Precipitated silica" could have been precipitated from any one of several solutions or from the action of water on silicon fluoride gas. The products may differ markedly for different types of precipitated silica due to minute amounts of retained impurities. In many accounts neither the source of silica is given nor the form in which it was used. The terms are given as they appear in the literature.

Other constituents present.—**H_2O:** Water has been a major constituent in most experiments, usually under confining pressure and at elevated temperature. Water at atmospheric pressure has been used at room temperature by Bergmann and Siegling (before 1845) and as superheated steam by Quensel (1906).

Concerning the system of H_2O-SiO_2 Morey and Fenner (1917, p. 1209), write

> "the solubility of SiO_2 in H_2O is very slight; so slight, in fact, that its amount is not enough to affect the ordinary fixed points in the binary system H_2O-SiO_2, such as the triple point, boiling point, or critical point."

However, more recent work by Ingerson and Morey (1940) has shown that at temperatures and pressures well above the critical point considerable amounts of silica may be carried by gaseous water.

Al_2O_3: Daubrée (1854); Friedel and Sarasin (1879; 1881); Chrustschoff (1891); Friedel (1891); Morozewicz (1893); Chrustschoff (1895); Schlaepfer and Niggli (1914); Müller and Königsberger (1906; 1918).

Alumina has been used in more experiments resulting in the formation of quartz crystals than has any other substance, except, of course, silica in the forms listed above. The number of experiments, however, may not constitute an accurate index of the importance of alumina, since in most cases the investigators were trying to produce feldspar, and quartz was formed as a by-product.

Na_2O or "soda": Friedel (1891); Chrustschoff (1895); Königsberger and Müller (1906); Morey and Bowen (1927); Bowen, Schairer, and Willems (1930); Kracek (1939).

Friedel (1891) used a "soda solution" on amorphous alumina to produce corundum and diaspore. When the alumina contained a little silica, quartz crystals were formed.

Chrustschoff (1895) reports "soda" as one of the eight constituents which he heated to 550° in a sealed platinum vessel and from which quartz, hornblende, diopside, orthoclase, and anhydrous silicate were reported as formed.

Na_2O is reported in the chemical analyses of the glass used by Königsberger and Müller (1906); Morey and Bowen (1927); Bowen, Schairer, and Willems (1930); and Kracek (1939).

Fe_2O_3: Chrustschoff (1891; 1895); Königsberger and Müller (1906); Morey and Bowen (1927); Bowen, Schairer, and Willems (1930).

Ferric oxide was one of eight constituents used by Chrustschoff to make hornblende in experiments in which quartz was a by-product. It was reported in the chemical analyses of the glass used by Königsberger and Müller (1906); Morey and Bowen (1927); and Bowen, Schairer, and Willems (1930).

CaO: Daubrée (1854); Königsberger and Müller (1906); Schlaepfer and Niggli (1914); Morey and Bowen (1927).

Daubrée produced crystallized quartz by the action of silicon chloride vapor on lime at red heat. CaO was reported in the chemical analyses of the glass used by Königsberger and Müller (1906); Schlaepfer and Niggli (1914); and Morey and Bowen (1927).

K_2O or "potash": Friedel and Sarasin (1879); Chrustschoff (1895); Müller and Königsberger (1918); Morey and Bowen (1927); Michel-Lévy and Wyart (1939).

"Potash" was reported as a constituent of the mixtures used by Friedel and Sarasin (1879) to make orthoclase and by Chrustschoff (1895) to make hornblende.

Michel-Lévy and Wyart (1939) produced quartz by detonation of vitreous silica with water and "potash" in a silver tube, but with vitreous silica and water alone the product was glass.

K_2O was reported in the chemical analyses of the glass used by Müller and Königsberger (1918); Morey and Bowen (1927).

KOH: Chrustschoff (1891); Schlaepfer and Niggli (1914).

A few drops of potassium hydroxide were added to the mixture from which Chrustschoff (1891) produced hornblende, analcime (?), adularia (?), and quartz crystals.

Schlaepfer and Niggli (1914) made orthoclase, hieratite, and quartz from a mixture of potassium hydroxide, water, and silica [3] in a silver-lined bomb.

[3] Precipitated from SiF_4.

NaOH: Maschke (1872); Chrustschoff (1891); Friedel (1896).

Maschke (1872) dissolved amorphous silica in a sodium hydroxide solution and heated the resulting clear solution in a sealed glass tube within a steel bomb. A small amount of a substance that "behaved the same way as crystalline quartz" was formed when the bomb was held at 175°-185° for 24 hours. At 195°-205° there was a marked increase in the amount of quartz, easily recognized in angular, birefringent grains.

Friedel (1896) investigated the action of sodium hydroxide solution on muscovite in a platinum-lined steel tube at temperatures around 500° C. If the solution contained enough silica so that the SiO_2:Na_2O was 4:1 an abundant deposit of quartz in bipyramidal crystals formed after 2 or 3 days.

Cu: Friedel and Sarasin (1879); Friedel (1891); Michel-Lévy and Wyart (1940).

Michel-Lévy and Wyart (1940) found that detonation of vitreous silica with water in a silver tube produced a glass, whereas the same experiment performed in a red copper tube resulted in the formation of quartz. They concluded that native copper was a mineralizing agent for quartz, at least between 450° and 465° C. This experiment supports the suggestion that copper should be considered as a constituent of any experiment performed in a copper-lined container. Quartz crystals were produced by Friedel and Sarasin (1879) and by Friedel (1891) in copper-lined bombs.

$CaCO_3$: Morozewicz (1893); Spezia (1899); Baur (1911).

Morozewicz (1893) made a mixture of the composition of liparite (Lipari rhyolite) using carbonates instead of oxides for the charge. When the mixture had been heated in a crucible at various temperatures for 31 days clouds of microscopic crystals were produced in clear, well-formed, double pyramids.

Spezia (1898) put sulfuric acid and a small piece of magnesite in the bottom of a platinum crucible. A small receptacle of silver above, which did not touch the acid, contained a solution of sodium silicate with a little powdered calcite. After 7 days' heating at 280°-290° C. in a steel bomb containng a small quantity of water, quartz crystals visible under low magnification had formed.

Baur (1911) produced quartz and pectolite by heating 7.8g. waterglass, 0.4g. $NaAlO_2$, 0.3g. SiO_2, 5.5g. H_2O at 450° in a steel bomb for 12-16 hours. The same minerals were produced when 7.8g. waterglass, 1.0g. $CaCO_3$, and 5.4g. H_2O were used.

KCl: Friedel and Sarasin (1883); Day and Shepherd (1906); Brun (1908).

Friedel and Sarasin (1883) produced many quartz crystals and no albite from a mixture of sodium silicate, aluminum silicate, and potassium chloride. When the experiment was repeated under apparently identical conditions large albite crystals formed. When potassium chloride was present in great excess, both orthoclase and albite resulted.

Day and Shepherd (1906b, p. 1099) "obtained quartz crystals from glass by the use of 80 per cent potassium chloride and 20 per cent lithium chloride at all temperatures below 760° C." Their experiments were performed in an iridium crucible in an iridium-lined furnace.

Brun (1908) exposed masses of fused silica to KCl and NaCl vapors (mixed) out of contact with air at 700°-750° C. for 3 or 4 days and produced quartz crystals 3 millimeters long.

NaCl: Spezia (1908) found that the addition of 12.7 per cent NaCl to a 1.9 per cent aqueous solution of Na_2SiO_3 resulted in the formation of clearer crystals with more marked trigonal symmetry and with striations on the prism faces parallel to the edge between the trapezohedron and the prism (the edge xm). According to Spezia solutions containing NaCl favored the growth of left-handed crystals.

LiCl: Hautefeuille and Margottet (1881); Lenarčič (1904); Day and Shepherd (1906).

Hautefeuille and Margottet (1881) produced doubly terminated prisms of quartz as well as tridymite and silicates of lithium by using lithium chloride as a flux at temperatures below the melting point of silver (960° C.). No description is given of the apparatus used or of the other constituents of the experiment.

Lenarčič (1904) added 1 gram of lithium chloride, 1 gram of magnesium fluoride, and 1 gram of molybdic acid to 5 grams of a mixture of the chemical composition of a mica schist. This mixture was heated, apparently above 1000° C., for about half a day, and plates of quartz formed which were visible in thin section.

Day and Shepherd's experiment with lithium and potassium chlorides has already been briefly described.

$Na_2WO_4 \cdot 2H_2O$: Hautefeuille (1878); Bauer (1899); Fenner (1913).

Hautefeuille (1878) heated a dry melt of amorphous silica with sodium or lithium tungstate at 800°-950° C. for several hundred hours and produced doubly terminated hexagonal pyramids.

Bauer (1899) thought that the presence of about 8 per cent sodium tungstate in a mixture of eight constituents was a significant factor in the formation of quartz from the dry melt. He cited the work of Hautefeuille (1878) and of Morozewicz (1893) who made quartz crystals in a dry melt containing 1 per cent tungstic acid.

Fenner (1913) chose sodium tungstate as a flux in his study of the stability relations of the silica minerals because it melts at a low temperature and does not volatilize at a high temperature.

CONCLUSIONS

Study of the literature indicates that quartz has been made under a variety of physical-chemical conditions on many occasions. The exact number of successful experiments is unknown, but it is certainly large. The most favorable conditions under which quartz may be crystallized are only broadly outlined, but certain temperature-pressure relationships appear to be more favorable than others, and certain forms of silica are perhaps more desirable sources. Several catalysts have been mentioned which may be effective, and phase-rule considerations have been pointed out which may ultimately prove essential.

If great intervals of geologic time are required for crystallization, laboratory experiments will probably always result in small crystals. Either time may be reduced by an increase in pressure, or protracted time intervals may not be as important a prerequisite as the duration of certain experiments suggests. In any event the possibility of producing crystals large enough for oscillator use artificially and in a reasonable length of time has not been disproven. Although much experimental work will be necessary before such crystallization can be brought about the record of experiments hitherto performed should not discourage the attempt.

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REFERENCES		SILICA													SILICATES					GLASS				
YEAR	OBSERVER	Chalcedony	Quartz	Tridymite	Cristobalite	Amorphous SiO_2	Dialyzed SiO_2	Gelatinous SiO_2	SiO_2 solution	Precipitated SiO_2	Fused SiO_2	$SiO_2 \cdot 3\ H_2O$	SiO_2	Water glass	Na-silicate	K-silicate	Al-silicate	K-Al-Silicate	Type unidentified	Lead glass	Manganese glass	Calcium glass		
....	Bergmann								X															
....	Siegling								X															
1845	Schafhäutl									X														
1849	Daubrée																							
1851	Sénarmont														X									
1851	"														X									
1854	Daubrée																							
1854	"																							
1854	"																							
1854	"																							
1862	Déville															X								
1872	Benrath					X																		
1872	Maschke					X																		
1872	"					X																		
1873	Chrustschoff						X																	
1878	Hautefeuille					X																		
1878	"																							
1879	Daubrée																							
1879	"																							
1879	"																							
1879	"																							
1879	"																							
1879	Friedel and Sarasin							X																
1880	Marsden										X													
1881	Chrustschoff						X																	
1881	"						X																	
1881	Friedel and Sarasin							X																
1881	Hautefeuille and Margottet												X											
1883	Friedel and Sarasin														X		X							
1887	Chrustschoff						X																	
1889	Bruhns																							
1889	"																							
1889	"												X											
1891	Chrustschoff						X																	
1891	Friedel												X											
1893	Morozewicz											X												
1895	Chrustschoff							X																
1895	"							X																
1895	"							X																
1895	"							X																
1895	"																							
1895	"																							
1896	Friedel												X											
1898	Spezia														X									
1898	"														X									
1898	"														X									
1899	Bauer																							
1902	Baur					X																		
1902	"					X																		
1904	Lenarčič																							
1904	"																							
1904	"																							
1904	"																							
1904	"																							
1905	Spezia		X												X									
1905	"		X												X									
1906	Allen, Wright, and Clement														X									
1906	Day and Shepherd																				X			
1906	Königsberger and Müller																				X			
1906	" "																				X			
1906	" "																							
1906	" "																							
1906	Quensel												X											
1906	"												X											
1906	"												X											
1906	"												X											
1906	"												X											
1907	Day, Allen, Shepherd, White, and Wright														X									
1908	Brun										X													
1908	Spezia		X												X									
1911	Baur												X											
1911	"												X	X										
1911	"												X	X										
1911	"													X										
1911	"												X											
1912	Fenner										X													
1912	"									X														
1912	"			X																				
1912	"				X																			
1913	Fenner										X													
1913	"			X																				
1913	"		X																					
1913	"								X															
1913	"	X																						
1914	Schlaepfer and Niggli												X											
1914	" "								X															
1914	" "								X															
1914	" "								X															
1917	Morey and Fenner								X							X								
1917	" "								X							X								
1917	" "								X							X								
1917	" "								X							X								
1917	" "								X							X								
1918	Müller and Königsberger															X								
1918	" "															X								
1918	Wilson																					X		
1918	"																				X			
1925	Weil								X						X									
1927	Ipatieff and Mouromtseff							X																
1927	Morey and Bowen																				X			
1927	" "																				X			
1927	Rao																							
1930	Bowen, Schairer, and Willems												X											
1936	Trostel									X														
1939	Kracek												X											
1939	Michel-Lévy and Wyart												X											
1940	Ingerson and Morey												X											
1940	Michel-Lévy and Wyart									X														
1940	"									X														
1941	Geophysical Laboratory		X										X		X									
1941	" "		X		X																			

TABLE 1.—*Summary of experiments resulting in production of quart*

| CONSTITUENTS EMPLOYED IN EXPERIMENTS | | |

...nts resulting in production of quartz

S	OXIDES												MINERALS								ROCKS						MISC.		TEMPERATURE (Centigrade)	TIME	PRESSURE Lbs. per sq. in.
	BeO	K_2O	Na_2O	Li_2O	CaO	MgO	$Al_2O_3(H=Al_2O_3\cdot 3H_2O)$	FeO	Fe_2O_3(H = hydrous)	CO_2	WO_3	H_2O	Orthoclase	Albite	Oligoclase	Muscovite	Hornblende	Pyroxene	Realgar	Orpiment	Basalt	Obsidian	Biotite gneiss	Chlorite schist	Mica schist	Hexogene	Mineral waters			
												X																room	2 years	15
												X																room	8 years	15
												X																120° ± ?	8 days	30± ?
												X																"white heat"		15
												X						X												
	X				X							X							X												
						X						X																			
							X					X																			
												X																" >170°"	"long"	
												X																"low"	24 hours	
												X																175°–185°	" "	
												X																195°–205°		
												X																250°	10–14 days	
												X																750°, 800°–950°	"several hundred hours"	
												X																" " "	" " "	
												X																400°	1 month	15000(?)
												X	X															"		"
												X						X										"		"
												X										X						"		"
												X																"		"
		X					X					X															X	"	2 days	"
		X					X					X																"below dark red"	38 hours	
												X																>960°	"a number of hours"	15
												X																250°	10–14 days	
												X																350°	" " "	
		X					X					X																	16–30 hours	
												X																<960°		
												X																		
												X																250°–320°	7–24 days	
												X																300°	50 hours	
												X																"	10 hours	
												X																"	" "	
							X	X	H			X																550° ±	3 months	
							X																					450°–535°	6–10 hours	
							H																					1060°–1430°	31 days	15
		X					X					X																300°–400°		
		X	X				X		X			X																550°		
												X																250°		
												X																360°		
												X																240°–300°	5 hours	1290
												X																310°–360°	" "	2835
												X				X												500° ?	2–3 days	
X												X																275°–300°		
												X																300°	4 days	
												X	X	X		X	X											230°–240°	7 days	
												X																1100°		15
												X																520°	4–6 hours	
												X																"	" "	
																			X							X		>1020°	12½ hours	15
																						X						"	15½ hours	"
																						X				X		"	14½ hours	"
																						X						"	"	"
																								X				"	"	"
												X																164°–338°	6 months	<2250
												X																165°–337°	3½ months	"
												X																400°–450°	3 days	
																												<760°		
									X			X																360°	60 hours	
												X																"	" "	
												X																"	" "	
												X									X							320°		
												X									X						X	420°	42 hours	
											X	X			X													800°–1000°	60 hours	15
											X	X			X													" "	120 hours	"
											X	X			X													" "	4 hours	"
											X	X			X													" "	10 hours	"
											X	X			X													" "	6 hours	"
												X																400°–450°	3 days	
												X																700°–750°	2–4 days	
												X																168°–340°	5 months	<2250
												X																450°	12–16 hours	
												X																"	" " "	
												X																"	" " "	
												X																"	" " "	
												X																"	" " "	
												X																400°–450°	2–3 days	
												X																" " "	" " "	
												X																" " "	" " "	
												X																" " "	" " "	
												X																863°–875°	11⅛ hours	15
												X																854°–864°	73½ hours	"
												X																865°–877°	24 hours	"
												X																350°–380°	22 hours	
												X																800°–850°		
												X																470°	1–2 days	
												X																"	" " "	
												X																"	" " "	
												X																"	" " "	
				X								X																420°		
						X						X																480°		
												X																500°		
												X																520°		
												X																600°		
									X			X																320°	36 hours	
									X			X																420–430°	27 hours	
												X																	5 days	15
												X																650°–750°	7–15 hours	"
												X																310°–320°	30–40 hours	3000
												X																300°	20 hours	
												X																550°	" "	
																				X								32°–52°	6 months	15
								X																				*See text*		"
																												1300°	4 hours	15
		X	X																									642°–870°		
												X														X		550°	6 days	42000
												X																600°–1050°		3000
												X														X		450°–650°	4–13 days	>42000
												X														X		540°–585°	" " "	" "
												X																350°		1500
												X																			"

PRESSURE Lbs. per sq. in.	SIZE		FORM OF QUARTZ PRODUCED	APPARATUS OR PROCEDURE
	Length mm.	Diameter mm.		
15			rhombohedrons	Autoclave ("Papin digester")
15			pris. and rhom.	Crucible and earthenware retort.
30± ?	microscopic		triangular faces	
15	very small			
			pris. and rhom.	Sealed glass tube in steel tube containing water
			" " "	" " " " " " " "
			" " "	Chloride vapor passed over solids.
			" " "	" " " " " "
			" " "	" " " " "
			" " "	" " " " "
	very small		granules	Sealed glass tube.
	microscopic		prisms	Devitrification.
	larger			Evacuated sealed glass tube in bomb.
	visible at 300×			" " " " " "
			pris. and rhom.	Sealed glass tube
			" " "	
			" "acute"	
15000(?)	2		pris. and rhom.	Sealed glass tube in sealed iron tube
"	1–2			" " " " " " " "
"	" "			" " " " " " " "
"	" "			" " " " " " " "
"				" " " " " " " "
	0.5	0.1	pris. and rhom.	Sealed steel tube, copper-lined
15			prisms	Silica crucible
	Visible at 300×		pris. and rhom.	Sealed glass tube
	" " "		hex. plates	Steel bomb, gold lined
	Measurable			Sealed steel tube, copper- or platinum-lined
			pris. and rhom.	
				Sealed platinum tube in sealed steel tube
	8	3	pris. and rhom.	Sealed glass spheres
				Sealed platinum tube in sealed steel tube
				" " " " " " " "
				" " " " " " " "
	0.2			Evacuated sealed glass tube in sealed iron tube
				Copper-lined steel tube in cast iron block
15	.005±		pris. and rhom.	Crucible
				Sealed platinum vessel
				" " "
				" " "
				" " "
				" " "
				" " "
1290				
2835			pris. and rhom.	Steel tube, platinum-lined
				Sealed steel tube
				Glass tube in steel tube
				Platinum and silver crucibles in steel bomb
15			irregular	Porcelain crucible
				Sealed steel tube
				" " "
15				Crucible
"			plates	"
"			"	"
"			"	"
"			"	"
<2250			pris. and rhom.	Silver-lined bomb
"			" " "	" " "
	20 added		doubly term. crystals	Platinum crucible in steel bomb
	3±			Iridium crucible in iridium furnace
	.01–.05	.004–.02	doubly term. crystals	Platinum-lined bomb containing filter
	0.6			" " " " " " " "
				" " " " " " " "
				" " " " " " " "
15	Visible at 800×			Crucible. Jet with superheated steam.
"				" " " " " " " "
"				" " " " " " " "
"	visible at 1000×			" " " " " " " "
"	.05			" " " " " " " "
	2		doubly term. crystals	Steel bomb
	3			
<2250	14 added		pris. and rhom.	Silver-lined bomb
				Steel bomb
				" "
				" "
				" "
				" "
				" "
				" "
				" "
15	Up to 0.1		doubly term. crystals	Platinum crucible over Bunsen flame
"	"		" " "	" " " " " "
"	"		" " "	" " " " " "
	small		"perfect"	Bomb
			poor	Silver crucible in silver-lined autoclave
				" " " " " " " "
				" " " " " " " "
				" " " " " " " "
				Gold crucible in bomb
				" " " " " "
				" " " " " "
				" " " " " "
	0.1			Platinum-lined bomb containing filter
	0.01			" " " " "
15			rhombohedrons	Glass furnace (collapsed)
"				" " "
3000			pris. and rhom.	Incompletely sealed tube
			plates	Silver tube in bomb with hydrogen
	"large"			Steel bomb
				" "
15				Open glass tube
"				In platinum foil in electric furnace
15	.0180			Crucible?
				In platinum foil in electric furnace
42000	0.4	0.1–0.2	pris. and rhom.	Bomb. Detonation
3000				Bomb
>42000				In copper tube in bomb
" "				" " " " "
1500	0.37/day			Bomb.
"	0.01/day			"